Modeling Stream Thermal Dynamics: The Influence of Beaver Dams in a Minnesota Watershed

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Thanks for all your encouragement!
Dedication

To my family:
for their love and support
and for encouraging me to ask good questions.
ABSTRACT

Beaver dams are known to alter the thermal regime of ponds, streams, and adjacent subsurface waters. Downstream of a dam, stream temperature is influenced by increased exchange with the hyporheic zone, which may cool and buffer the stream’s diel temperature cycles. Concurrently, reduced shading in the beaver forage zone is likely to increase heat flux at the stream-atmosphere boundary. The dynamics of these processes can be analyzed to understand how stream temperature is affected on diel time scales, as well as longitudinally at distances downstream from the dam. At two beaver dam-impacted stream sites in the Knife River Watershed in Minnesota, USA, I monitored in-stream and shallow subsurface flow and temperature during low-flow summer conditions. I used a dye tracer test, vertical heat transport modelling, and soil characterization to estimate flux through the streambed at multiple locations. Temperature, stream flow, and atmospheric data were also collected throughout the summer from the two sites. A one-dimensional model of longitudinal stream temperature, calibrated to in-stream temperature measurements, was developed to determine which physical parameters and heat flux components have the greatest influence on stream temperature. The model was then used to demonstrate how these changes persist downstream, as well as to simulate stream temperature under potential future site conditions. These findings increase scientific understanding of stream temperature regime in the context of beaver dam-altered watersheds.
# Table of Contents

List of Tables ........................................................................................................................................ vi
List of Figures ......................................................................................................................................... vii
List of Notations .................................................................................................................................. viii
I.  INTRODUCTION ................................................................................................................................. 1
II.  LITERATURE REVIEW ...................................................................................................................... 3
    1.  Beaver-altered hydrology ................................................................................................................. 3
    2.  Beaver-altered thermal regimes ....................................................................................................... 4
    3.  Beaver impacts to aquatic ecosystems ......................................................................................... 5
    4.  Heat exchange at stream-atmosphere interface .......................................................................... 6
    5.  Heat exchange at streambed interface ........................................................................................ 7
    6.  Stream temperature methods and modeling ........................................................................... 8
III. SITE DESCRIPTION .......................................................................................................................... 9
    1.  Site 2B overview ............................................................................................................................. 9
    2.  Site 4B overview .......................................................................................................................... 13
IV.  DATA COLLECTION METHODS .................................................................................................... 15
    1.  Monitoring equipment installation ............................................................................................... 15
    2.  Site soil characterization ............................................................................................................ 18
    3.  Stream data .................................................................................................................................. 20
       i.  Temperature ............................................................................................................................. 20
       ii. Velocity and Discharge ........................................................................................................... 20
    4.  Tracer testing ............................................................................................................................ 21
       i.  Tracer testing layout and conditions .................................................................................... 21
       ii. Tracer testing equipment ........................................................................................................ 22
       iii. Tracer testing timeline .......................................................................................................... 23
       iv. Laboratory testing of tracer samples .................................................................................. 23
       v.  Hydraulic conductivity values .............................................................................................. 24
V.   HEAT FLUX MODELING METHODS .............................................................................................. 25
    1.  Stream surface heat and mass flux .............................................................................................. 26
       i.  Radiative heat flux ..................................................................................................................... 26
       ii. Evaporative heat flux ............................................................................................................... 29
       iii. Sensible heat flux .................................................................................................................... 30
2. Streambed heat and mass flux ................................................................. 30
   i. Advective heat flux ................................................................. 30
   ii. Conductive heat flux ............................................................. 32
3. Longitudinal modeling ................................................................. 32
   i. Model calibration ............................................................... 34
   ii. Sensitivity analysis ............................................................. 35
VI. RESULTS .................................................................................. 35
1. Tracer testing ................................................................. 35
2. Soil characterization ............................................................. 37
3. Stream and air temperature ................................................... 39
   i. Diel temperature cycles ....................................................... 39
   ii. Longitudinal stream temperature .................................... 41
4. Heat flux to the stream: atmospheric processes ..................... 43
5. Heat flux to the stream: streambed processes ....................... 45
6. Total heat budget ................................................................. 50
7. Model results ................................................................. 54
8. Model sensitivity analysis .................................................... 57
9. Model validation ................................................................. 57
VII. DISCUSSION ........................................................................... 63
1. Discharge ................................................................. 64
2. The stream surface boundary ........................................... 65
3. The streambed boundary .................................................... 66
4. Hydraulic gradient ............................................................. 67
5. Spatial resolution ................................................................. 67
6. Stream geomorphic units .................................................... 68
7. Aquatic habitat ................................................................. 68
8. Modeling ................................................................. 70
9. Future Work ................................................................. 70
VIII. CONCLUSION ......................................................................... 70
IX. BIBLIOGRAPHY .......................................................................... 72
X. APPENDIX .................................................................................. 84
List of Tables

Table 1. Monitoring locations at Site 2B .................................................................11
Table 2. Monitoring locations at Site 4B .................................................................14
Table 3. Phase one equipment installation .............................................................16
Table 4. Phase two equipment installation .............................................................17
Table 5. Hydraulic conductivity calculated from tracer testing ..............................37
Table 6. Soil texture estimated from soil particle analysis ........................................38
Table 7. Soil properties estimated from bulk density analysis ..................................38
Table 8. Sensible and evaporative heat flux ..........................................................45
Table 9. Site 2B and 4B pond bed and streambed seepage ......................................47
Table 10. Conductive and advective heat flux .......................................................50
Table 11. Net stream heat flux .............................................................................51
Table 12. Absolute total stream heat flux ..............................................................53
Table 13. Statistical RMSE and NSE model calculations .........................................57
Table 14. Coordinate locations and elevation data ...............................................84
Table 15. Site 2B maximum daily temperature ......................................................85
Table 16. Site 4B maximum daily temperature ......................................................86
Table 17. Equipment specifications ...................................................................87
List of Figures

Figure 1. Site 2B and Site 4B location.................................................................10
Figure 2. Surficial geologic formations...............................................................11
Figure 3. Impoundment at 2B-M-pond...............................................................12
Figure 4. Impoundment at 4B-U-pond...............................................................13
Figure 5. Impoundment at 4B-L-pond...............................................................15
Figure 6. Study timeline..................................................................................16
Figure 7. Monitoring equipment......................................................................18
Figure 8. Diagram of heat fluxes to the stream cross section..............................24
Figure 9. Diagram of stream shading parallel to the solar azimuth......................27
Figure 10. Diagram of stream shading normal to the cross-sectional width..........28
Figure 11. Fluorimeter measurements from tracer testing..................................36
Figure 12. Air temperature during the study period duration............................39
Figure 13. Hourly average air and stream temperature....................................40
Figure 14. Pond bed and pond surface temperature at Site 4B and Site 2B............41
Figure 15. Site 2B longitudinal temperature....................................................42
Figure 16. Site 4B longitudinal temperature....................................................43
Figure 17. Heat flux from solar radiation..........................................................44
Figure 18. Sediment temperature gradient and advection rate...........................46
Figure 19. Pond and streambed advection at Site 2B..........................................48
Figure 20. Pond and streambed advection at Site 4B..........................................49
Figure 21. Average daily heat fluxes at Site 2B..................................................52
Figure 22. Average daily heat fluxes at Site 4B..................................................52
Figure 23. Simulated and observed stream temperature at 2B-M-DS-1................54
Figure 24. Simulated and observed stream temperature at 2B-M-DS-2................55
Figure 25. Simulated and observed stream temperature at 4B-U-DS-1................56
Figure 26. Simulated and observed stream temperature at 4B-L-DS-1................56
Figure 27. Simulation of 4B-U-DS-1 temp in response to varying discharge...........58
Figure 28. Simulation of 4B-U-DS-1 temp in response to varying upstream temperature........58
Figure 29. Simulation of 4B-U-DS-1 temp in response to varying air temperature......59
Figure 30. Simulation of 4B-U-DS-1 temp in response to varying hydraulic conductivity........60
Figure 31. Simulation of 4B-U-DS-1 temp in response to varying vegetation height.....60
Figure 32. Simulation of 4B-U-DS-1 temp in response to varying net radiation .......61
Figure 33. Validation of the 4B-U-DS-1 model with average streambed heat flux.........62
Figure 34. Validation of the 4B-U-DS-1 model with hourly average streambed heat flux ....63
# List of Notations

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Name</th>
</tr>
</thead>
<tbody>
<tr>
<td>$A$</td>
<td>Cross sectional area (m$^2$)</td>
</tr>
<tr>
<td>$\alpha$</td>
<td>Vertical heat coefficient</td>
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<tr>
<td>$A_r$</td>
<td>Temperature amplitude ratio</td>
</tr>
<tr>
<td>$B_d$</td>
<td>Bank setback distance (m)</td>
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<tr>
<td>$B_h$</td>
<td>Streambank height above the water (m)</td>
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<tr>
<td>$B_r$</td>
<td>Bowen ratio</td>
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<td>Average heat capacity of soil solids (J/kg °C)</td>
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<td>Volumetric heat capacity of water (4.18×10$^6$ J/m$^3$ °C)</td>
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<td>Specific heat of water (4182 J/kg °C)</td>
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<td>$P_t$</td>
<td>Period of temperature variation</td>
</tr>
<tr>
<td>$\phi$</td>
<td>Porosity</td>
</tr>
<tr>
<td>$Q$</td>
<td>Stream discharge (m$^3$/s)</td>
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</table>
$q$  Darcy flux (m/s)
$q_v$  Streambed seepage (m/s)
$RH$  Relative humidity
$RMSE$  Root mean square error
$\rho_b$  Bulk density (kg/m$^3$)
$\rho_s$  Particle density (kg/m$^3$)
$\rho_w$  Density of water (1000 kg/m$^3$)
$S$  Stream gradient (m/m)
$S_a$  Atmospheric heat flux (W/m$^2$)
$S_b$  Streambed heat flux (W/m$^2$).
$S_{ab}$  Solar radiation absorbed at the water surface (0-1)
$S_al$  Solar altitude (radians)
$S_{as}$  Stream aspect (radians)
$S_{az}$  Solar azimuth (radians)
$S_w$  Solar radiation that reaches the water surface (0-1)
$\sigma$  Stefan Boltzmann constant ($5.67 \times 10^{-8}$ W/m$^2$ °C$^4$)
$T_a$  Air temperature (°C)
$T_l$  Temperature of the lower sensor (°C)
$t_i$  Initial time of the tracer test (s)
$T_{max}$  Daily maximum temperature (°C)
$T_{min}$  Daily minimum temperature (°C)
$T_o$  Observed stream temperature (°C)
$T_p$  Predicted stream temperature (°C)
$t_p$  Time to peak of the dye plume (s)
$t_t$  Travel time (s)
$T_u$  Temperature of the upper sensor (°C)
$T_w$  Water temperature (°C)
$\theta_m$  Gravimetric water content (kg/kg)
$\theta_v$  Volumetric water content (m$^3$/m$^3$)
$U$  Windspeed (m/s)
$V$  Volume of the sample collection tin (m$^3$)
$v$  Velocity of the thermal front (m/s)
$V_h$  Vegetation height above the water (m)
$v_s$  Seepage velocity (m/s)
$W$  Width (m)
$W_n$  Shadow length normal to the stream (m)
$W_p$  Shadow length parallel to the solar azimuth (m)
$x$  Longitudinal stream distance (m)
$\Delta z$  Difference in depth between temperature sensors (m)
I. INTRODUCTION

Beavers impact the hydrology and morphology of a watershed in a variety of ways. By building dams that retain water, beavers slow stream discharge and attenuate peak flows (Puttock et al., 2016). Although surface water stagnation is known to increase evaporation rates, this loss may be offset by decreases in peak-flow discharge (Woo and Waddington, 1990). Further, beaver dams alter flow pathways and connectivity between surface, hyporheic, and groundwater reservoirs (Westbrook et al., 2006). Slowed surface water, particularly in ponds, is more likely to seep into the subsurface – thus increasing hyporheic storage and groundwater recharge (Majerova et al., 2015). Increased water storage, in turn, raises the water table and may result in greater stream discharge rates under dry, low-flow conditions (Nyssen et al., 2011).

Beaver dam-induced hydrologic changes to ponds, streams, and adjacent subsurface zones are accompanied by changes to the thermal regime. Water temperature is controlled by upstream temperature, heat exchange at the water-atmosphere interface, and heat exchange at the water-bed interface (Sinokrot and Stefan, 1993; Poole and Berman, 2001). When a pond forms above a beaver dam, the increased water surface area allows for greater heat exchange at this boundary (Johnston and Naiman, 1987). Additionally, beaver foraging reduces vegetative cover, increasing direct solar radiation to the pond and stream (Johnston and Naiman, 1987; Johnston and Naiman, 1990). Thus, atmospheric variables such as air temperature and solar radiation may strongly influence surface water temperature near impoundments (Majerova et al., 2015).

Beaver dams are also known to increase water heat and mass flux at the streambed boundary (Johnson, 2004; Woo and Waddington, 1990). When a stream is impounded, pond elevation increases hydraulic head gradient and thus downwelling through the pond bed (Lautz and Siegel, 2010). Infiltrated water then follows a curvilinear pathway through the shallow subsurface, upwelling downstream of the beaver dam (Hester et al., 2009). Because the hyporheic zone increases residence time, decreases exposure to solar radiation, and dissipates heat through the subsurface sediments, hyporheic exchange is known to cool and buffer stream temperature during warm summer periods (Arrigoni et al., 2008). Accordingly, increased hyporheic flow from beaver impoundments may be an important process to consider in regions where high stream temperature is a concern.
Beaver hydrology and thermal regime may significantly influence the productivity and health of an aquatic ecosystem (Collen and Gibson, 2001; Gibson and Olden, 2014). In cold-water stream studies (based primary in the western U.S.), beaver impoundments have been found to improve thermal stability and thus increase species abundance (Bryant, 1983). In the Great Lakes region, where cold-water streams are often marginal, there is concern that beaver impoundments could raise water temperatures beyond the tolerable range for salmonid populations (Patterson, 1951; Peterson, 2012). Previous studies in Minnesota, Wisconsin, and Michigan have found variable thermal impacts associated with beaver dam presence or removal (Johnson-Bice et al., 2018).

In this study, I have analyzed flow and temperature patterns along two headwater tributaries in the Knife River watershed (KRW), a 218 km² watershed on the North Shore of Lake Superior in Minnesota, USA. The Knife River has a flashy hydrogeomorphic regime, characterized by high flows during the spring (average monthly discharge is 9.9 m³/s in April) and low flows by late summer (average monthly discharge is 0.9 m³/s in August) (USGS, 2020). Because the Knife River is a designated trout fishery, a significant effort is being made to protect its cold-water habitat. Beaver dams in this region are thought to raise stream temperature, and the Minnesota Department of Natural Resources (Minnesota DNR) removes dams from the main branches of the Knife River each spring (Goldsworthy et al., 2016).

Stream temperature in the KRW has historically been studied in the context of trout management, with mixed results (Smith and Moyle 1944; Hale, 1966). The Minnesota DNR recently measured stream temperature at four beaver dam sites, and found one instance of significantly elevated downstream temperature; the other three sites had similar temperatures above and below the dam (Peterson, 2012). As these studies indicate, the processes that affect stream temperature in the KRW may be complex. However, without high resolution measurements and physical context, it is difficult to understand which heat transfer processes are important at different spatial and time scales.

During July and August 2019, I measured site conditions – including topography, surficial geology, discharge, and stream temperature - at two beaver-impounded tributaries in the KRW. I used these data to answer the following questions:
1. To what extent are stream heat and mass flux dynamics altered by beaver dam presence?
2. How do these fluxes vary longitudinally and at different timescales?
3. How do physical topography and subsurface characteristics mediate stream heat and temperature regime?

In my approach, I examined both sites using a reach-scale analysis that encompassed the beaver pond as well as the stream segment above and below the impoundment. Differences between the two sites – including stream gradient, amount of vegetative shading, and substrate material – were also examined in this study.

II. LITERATURE REVIEW

The North American beaver (Castor canadensis) is considered a keystone species because of its ability to drastically modify the hydrology, ecology, and biochemistry of wetland ecosystems (Naiman et al., 1986; Mills and Doak, 1993). Beavers primarily accomplish this ecosystem alteration through the construction of composite dams that impound stream water (Gurnell, 1998). The resulting ponds and channels provide shelter, expand feeding areas, and allow beavers to avoid predators (Zurowski, 1992). The dams, in turn, alter stream discharge, water temperature, flow pathways, and water storage capacity (Rosell et al., 2005; Majerova et al., 2015). Beaver impacts to hydrology are mediated by features of the physical landscape, including surficial geology, morphology, and characteristics of the dam itself. Disentangling and clarifying the relationship between these variables has been the focus of several studies (Gurnell, 1998; Levine and Meyer, 2014; Janzen and Westbrook, 2011).

1. Beaver-altered hydrology

The age and condition of a beaver impoundment – and accordingly, its hydrologic impacts – are dynamic through time (Polvi and Wohl, 2012). Actively-maintained beaver dams, consisting of tightly packed wood, debris, and sediment, have been found to retain large volumes of water and sediment (Puttock et al., 2016). Over time, and particularly once an impoundment has been abandoned, erosion of the dam sediments may result in greater permeability and through-flow (Woo and Waddington, 1990). As water depth decreases in the pond, vegetation is able to reestablish and cover progressively more of the surface area (Collen and Gibson, 2001).
Alternatively, high flows at any point in time may breach the dam, releasing built-up sediment and water (Levine and Meyer, 2014).

While high-gradient stream environments support relatively small beaver ponds, beaver dams in low-gradient streams may impound large, shallow pond areas (Johnston and Naiman, 1987). The extensive contact created between the water and pond bed in these impoundments, in tangent with long water residence times, has been found to enhance downwelling seepage through the pond bed (Lautz and Siegel, 2010). While some seepage water is thought to recharge the underlying aquifer, studies have found that beaver ponds increase flow to the downstream reach through hyporheic paths (Janzen and Westbrook, 2011). Flow attenuation has also been shown to increase pond and subsurface storage, resulting in greater discharge during dry periods (Gurnell, 1998; Burns and McDonnell, 1998). Work by Nyssen et al. (2011) suggests that a series of beaver ponds may enhance flow attenuation; successive beaver impoundments were found to delay peak flows and lengthen the recurrence interval of flood events.

2. Beaver-altered thermal regimes

Beaver dams alter the thermal regime of ponds, streams, and adjacent subsurface waters. Beaver impoundments effectively a) increase water surface area, thereby increasing atmosphere-surface water heat exchange (Cook, 1940), b) remove riparian vegetation, increasing direct solar radiation to the water surface (Beschta, 1987), and c) raise hydraulic head relative to downstream areas, forming a gradient that allows for greater infiltration to the subsurface (Hester and Doyle, 2008). In beaver ponds, heat exchange with the atmosphere may result in thermal stratification (Bonar and Petre, 2015). While this effect is significant during winter months, most beaver ponds are well-mixed and show minimal stratification during other seasons (Majerova et al., 2017). In most beaver ponds, depth, bed material, and exposure to solar radiation have a strong influence on water temperature (Johnson, 2004).

Immediately downstream of beaver ponds, in-stream temperature also reflects climate to a large extent. However, the influence of inputs from beaver ponds and subsurface storage adds complexity to understanding the thermal regime. Several researchers have observed increased temperatures downstream of beaver dams, while others have found decreased downstream temperatures (Kemp et al., 2012). While these results may reflect real variation, incongruity may also arise from low temporal/spatial resolution of measurements, or a failure to account for the
physical heterogeneity of the system. Majerova et al. (2017) found net warming of 1.15 °C at the beaver dam complex scale, but greater temperature variation in individual geomorphic units. In a study by Weber et al. (2017), beaver dams were found to have a buffering effect of stream temperatures, particularly during the summer.

3. Beaver impacts to aquatic ecosystems

Beaver hydrology and thermal regime directly influence the biodiversity and productivity of aquatic ecosystems (Pollock et al., 2003). These impacts to stream ecology have been widely studied, often for salmonid management purposes (Smith and Moyle 1944; Peterson, 2012). In cold-water streams where temperature is frequently below the optimal range for salmonid species, warming in beaver ponds may improve growth outcomes (Rutherford, 1955). In marginal cold-water habitat, however, warming has been associated with thermal stress and lower productivity (Knudsen, 1962). Beaver-induced thermal and sedimentation dynamics also impact the abundance of macroinvertebrates, which are an important food source for fish. While certain macroinvertebrate species may decline in beaver ponds, increased biomass (associated with increased littoral surface area) has been observed (Sprules, 1941; McDowell and Naiman 1986). In addition, beaver impoundments have been found to increase geomorphic heterogeneity and deep pool habitat, which provide thermal refuge to salmonids (Tate et al., 2007; Bouwes et al., 2018). Furthermore, beaver impoundments may enhance late-summer discharge rates and water storage (Tappe, 1942), which has been associated with greater salmonid productivity and habitat availability (Leidholt-Bruner et al., 1992).

In the Great Lakes region, potential warming poses a threat to already marginal cold-water trout streams. Climate change-driven warming is expected to decrease overall fish productivity, and cold-water species may be most adversely impacted (Huff and Thomas, 2014). There is also concern that beaver impoundments could exacerbate stream warming trends. However, a wide range of beaver thermal impacts have been documented in the Great Lakes region (Johnson-Bice et al., 2018). In low-gradient Wisconsin streams, Avery (1983, 2002) observed cooler temperatures in the years following dam removal. Meanwhile, in Michigan and western New York, in-stream features (including beaver dams) were found to promote hyporheic exchange and cooler downstream temperatures (White et al., 1987; Feiner and Lowry, 2015). Still other studies in Wisconsin have suggested that beaver dams do not significantly affect downstream
temperature (DuBois and Schram, 1993). The influence of beaver-altered thermal regimes on fish productivity in the Great Lakes region, while implied in several stream temperature studies, has been less quantitatively documented (Johnson-Bice et al., 2018).

### 4. Heat exchange at stream-atmosphere interface

Stream temperature is controlled by upstream source water temperature as well as exchange at the stream surface and streambed (Sinokrot et al., 1995). At the stream-atmosphere interface, heat exchange occurs via sensible heat transfer (as a result of air temperature / stream temperature differences), short-wave (solar) radiation, long-wave radiation, and evaporative flux (Caissie, 2006). Of these components, solar radiation has the greatest influence on stream temperature (Sinokrot and Stefan, 1993; Johnson, 2004). Solar radiation is also the primary driver of diel cycles in which stream temperature decreases at night and increases during the day.

In watershed studies, maximum daily stream temperature has been shown to decrease with increased shading from solar radiation (Johnson, 2004). Likewise, vegetation removal increases average monthly maximum stream temperature (Brown and Krygier, 1970). Beaver foraging is known to decrease vegetation cover around impoundments, which may increase direct radiation to the pond surface. Impoundments also increase water surface area, further driving heat flux at the water-atmosphere interface (Johnston and Naiman, 1987).

Air temperature (via evaporation and sensible heat transfer) also has an impact on stream temperature. Particularly in streams with small baseflow discharge, air temperature and stream temperature fluctuations may be closely corelated (Arismendi et al., 2014). Regression models (Mohseni and Stefan, 1999) and stochastic models (Caissie et al., 1998) that rely on climate variables, including air temperature, have thus been developed. In Minnesota, linear regression models have been developed to predict stream temperature under future climate change scenarios (Pilgrim et al., 1998). While the relationship between air and stream temperature may be linear under moderate conditions, some studies use logistic regression to account for evaporative cooling at high air temperature (Mohseni and Stefan, 1999).
5. Heat exchange at streambed interface

While heat flux at the streambed boundary has not been well-studied historically, in recent years it has been the focus of several studies (Gerecht et al., 2011; Caissie and Luce, 2017). Heat flux at the streambed is primarily a function of a) advection-driven mass flux through the boundary and b) conduction driven by differences in water and streambed temperatures (Hester et al., 2009). The relative importance of these processes at the streambed interface is mediated by physical streambed properties, channel geometry, and dynamics of the hyporheic zone. In some systems, approximately 15% of energy flux has been found to take place at the streambed boundary (Evans and Petts, 1998). In other cases, variation in flux at the streambed may be a minor component of the stream energy budget (Sinokrot and Stefan, 1993). Nonetheless, inputs from the hyporheic zone can represent an important buffer to stream temperature, particularly in streams with marginal habitat (Tague et al., 2007).

Downwelling to the hyporheic zone occurs when gradients in the streambed cause in-stream water to infiltrate into the shallow subsurface (Boano et al., 2014). Although advection-driven hyporheic exchange occurs in most streams on a small scale, the presence of in-stream topography or features may significantly affect the flow pathways and timescales of this process. Hyporheic exchange is enhanced by in-stream structures (e.g., debris dams, log dams, and boulder weirs) that cause a drop in hydraulic head (Kasahara and Wondzell 2003; Hester and Doyle, 2008). A drop in head induces curvilinear pathways that extend downwards beneath the structure and then upwards at a distance downstream (Thibodeaux and Boyle, 1987; Gooseff et al., 2006). Larger structures with larger head drops may induce greater hyporheic flux, although this relationship is non-linear and limited by depth to bedrock or impermeable sediments (Kasahara and Wondzell, 2003; Hester and Doyle, 2008).

Studies have found that in-stream structures induce increased hyporheic flow through underlying sediments (Lautz et al., 2006), and in doing so impact downstream temperatures. During the summer, relatively warm stream water is cooled due to heat exchange with the substrate and shielding from solar radiation as it travels through the hyporheic zone. When the water upwells to rejoin the surface stream, it may then buffer or cool surface stream temperatures (Arrigoni, 2008). Because beaver dams create pressure gradients that force subsurface water through longer and deeper pathways, they may have a significant influence on stream temperature (Beschta, 1987;
Keery et al., 2007; Majerova, 2017). This is apparent under low-flow conditions, when baseflow inputs are responsible for a greater portion of stream flow.

6. Stream temperature methods and modeling

In studies of the hyporheic zone, tracer testing has been used to understand how dams or other in-stream geomorphic structures (IGSs) affect subsurface flow velocity, dispersion, and travel time (Briggs et al., 2013). In a study by Hester et al. (2009), NaCl salt was injected into a piezometer upstream of a weir and conductance was measured in downstream piezometers to determine hyporheic velocity and travel time. Other studies have similarly used salt injections to determine hyporheic travel time before and after the removal of large woody debris (Wondzell, 2006; Wondzell et al., 2009). Additionally, both Rhodamine WT and NaCl were used by Majerova et al. (2015) in a study of beaver dam impacts on hyporheic and stream temperature regimes. In several experiments, hyporheic testing has also been used to characterize subsurface hydraulic conductivity (Kalbus et al., 2006).

Longitudinal stream temperature models – both statistical and deterministic - have been used to evaluate the relative importance of various stream parameters (Becker et al., 2004). While statistical models are less data-intensive to implement, deterministic models have the advantage that they resolve the stream heat budget and may therefore be used to simulate potential stream temperature scenarios (Dugdale et al., 2017). MNSTREM is an example of a well-established deterministic model that simulates stream temperature as a function of hydrogeologic and atmospheric data (Sinokrot and Stefan, 1993). The MNSTREM model solves the basic one-dimensional heat advection-dispersion equation

\[ A \frac{\partial T_w}{\partial t} + \frac{\partial (QT_w)}{\partial x} = AD_t \frac{\partial^2 T_w}{\partial x^2} + \frac{W(S_a + S_b)}{\rho_w c_p} \]  

Eq. 1

where \( T_w \) is stream temperature (°C), \( x \) is longitudinal stream distance (m), \( D_t \) is the dispersion coefficient (m²/s), \( A \) is cross sectional area (m²), \( W \) is width (m), \( Q \) is stream discharge (m³/s), \( c_p \) is specific heat of water (J/kg °C), \( \rho_w \) is the density of water (kg/m³), \( S_a \) is atmospheric heat flux (W/m²), and \( S_b \) is streambed heat flux (W/m²). MNSTREM can be used to simulate temperature on an hourly timescale, making it well suited to evaluating diel stream temperature fluctuations.

The Stream Network TEMPerature model/Stream Segment TEMPerature Model (SNTEMP/SSTEMP) is another one-dimensional deterministic heat transport model (Bartholow,
Although SNTEMP has a coarser time resolution – it is designed to estimate stream temperature on a daily scale using averaged values – it is capable of linking inflowing tributaries to larger streams, allowing for greater stream complexity. Studies using SNTEMP have further demonstrated how such models can be used to simulate changing site conditions, including vegetation removal (Sullivan et al., 1990) and channel restoration (Bartholow, 1993).

Beaver-impacted streams present the opportunity to apply stream temperature modeling in an environment where physical site conditions may be altered dramatically on short time scales. Downstream of a beaver impoundment, streams experience variable levels of solar radiation as shading vegetation is foraged and then regrows. The state of the dam itself is important to consider, because dam permeability can have a large impact on discharge and stream temperature immediately downstream. Hyporheic flow is also known to be induced by steep head gradients (Hester and Doyle, 2008), such as those between a beaver dam and the downstream reach.

Physical changes to the beaver impoundment and surrounding riparian zone – either due to the passage of time or a management decision – may be simulated to gain an understanding of how these changes would impact stream temperature.

III. SITE DESCRIPTION

During the summer of 2019, two tributaries in the Knife River Watershed (KRW) were selected for in-depth observation and analysis. These tributaries – Site 2B and Site 4B - each feature at least one intact beaver impoundment and are located in the forested headwaters of the watershed. From the beaver pond at each site, seepage through the dam and shallow subsurface feeds an outflowing downstream channel. Due to their proximity (7 km apart) Site 2B and Site 4B have comparable meteorological regimes. The KRW normal average precipitation in July is 9.83 cm, and normal average temperature is 18.4 °C (NOAA, 2020). However, Site 2B and Site 4B are distinct in their topography and surficial geology. Monitoring was conducted at multiple locations along the tributary at Site 2B and Site 4B to assess the stream and physical site conditions.

1. Site 2B overview

Site 2B is located in the hummocky upper region of the Knife River watershed (Figures 1 and 2). This region is low gradient, and numerous small ponds and headwater tributaries form here before flowing southeast towards Lake Superior. The site’s surficial geology consists of clay-rich
glacial tills overlain by loamy soils (Hobbs and Goebel, 1982). The till is moderately to highly impermeable, and potential groundwater recharge rate in the region is approximately 17 cm per year (Smith and Westenbroek, 2015).

Figure 1. Location of the main study area for Site 2B and Site 4B in the Knife River Watershed (LiDAR Elevation, 2011).

The glacial till at Site 2B is primarily from the Cromwell Formation, which was deposited by the Superior Lobe of the Wisconsin glaciation (Figure 2). The Superior Lobe advanced from the northeast to the southwest, reaching its maximum extent approximately 20,500 years ago (Wright et al., 1973). As it retreated, glacial diamicton was deposited across the landscape (Jennings and Johnson, 2011). In the region of the study site, drift thickness above bedrock is approximately 7 m (Jirsa, 2016).
Figure 2. Site locations and superficial geologic formations (Jirsa, 2016) in the Knife River Watershed.

At Site 2B, one beaver pond and the channel <100 m downstream was identified as the focal area for the study. All stream locations at Site 2B are numbered in order of their longitudinal distance downstream from this main pond (2B-M-pond) monitoring location (Table 1).

Table 1. Monitoring locations at Site 2B listed by longitudinal distance downstream from the 2B-M-pond monitoring location. The locations in the box are the main study locations.

<table>
<thead>
<tr>
<th>Location name</th>
<th>Distance downstream (m)</th>
<th>Depth to bedrock (m)*</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-US-1</td>
<td>-157</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-pond</td>
<td>0</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>11</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>37</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>63</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-4</td>
<td>110</td>
<td>6</td>
</tr>
<tr>
<td>2B-M-DS-5</td>
<td>300</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-6</td>
<td>327</td>
<td>7</td>
</tr>
<tr>
<td>2B-M-DS-7</td>
<td>340</td>
<td>7</td>
</tr>
</tbody>
</table>

*Jirsa, 2016
2B-M-pond is the farthest downstream of a long chain of beaver impoundments. At the time of the study, this pond had a surface area of 3250 m² and an average depth of 0.8 m. Aerial imagery suggests that the 2B-M-pond dam is at least seventeen years old (Google Earth, 2003), and that beavers have been active in this tributary for at least ten years prior to its construction (Google Earth, 1992). The dam is 45 m long and approximately 1.4 m tall, as measured on the downstream side. Although the dam is not being actively maintained, compacted sediment throughout the dam prevented leakiness and maintained consistent water levels throughout the period of observation. One point near the center of the dam contains less soil and more visible branches; water seeps through the base of this section to form a small outflowing channel (Figure 3).

![Figure 3. The 2B-M-pond dam (left) and pond (right) on May 14th, 2019.](image)

The outflowing stream flows through highly saturated beaver meadow - primarily riparian reeds and grasses - meandering and deepening with distance. The first monitoring location was in a shallow riffle 11 m downstream (slope = 7.8%). By the second monitoring location (37 m downstream, slope = 1.0%), the channel is nearly 0.7 m deep. By this point the channel is also shaded by dense riparian vegetation. Farther along, as the stream flows under forest cover, the channel bed transitions from muddy sand-loam to a higher proportion of gravel. The third downstream point, 2B-M-DS-3 (63 m ds, slope = 2.5%), was located in this relatively shallow segment. The meandering stream channel continues under forest cover for the next several hundred meters; the final four temperature sensors were in this shaded part of the reach.
2. Site 4B overview

Closer to Lake Superior, the glacial tills of the Cromwell formation are overlain by younger tills and glaciolacustrine sediments of the Barnum formation (Leverett, 1928; Jirsa, 2016). Site 4B is situated in this Superior Lobe Clay Plain region of the watershed (Figure 1). The region is characterized by its high proportion of low-permeability clays, which may impede groundwater recharge. However, due to landscape heterogeneity, average potential groundwater recharge reaches 21 cm per year (Smith and Westenbroek, 2015). Stream gradients in this geomorphic area are generally steeper, and slopes may be as high as 6-8% in some places (Jirsa, 2016). Stream drainage networks are often deeply incised in the clay soils; depth to bedrock may be as low as 2 m in the upstream (high elevation) regions of the Clay Plain. In farther downstream (low elevation) regions of the Clay Plain, stream incision to bedrock is common.

Site 4B spans this change in elevation and the transition from alluvial to bedrock-controlled channel. The main study location for Site 4B was more upstream (elevation = 258 m, slope = 2%), in a loamy clay and gravel-bedded reach. As with Site 2B, the study area for Site 4B was defined by a large pond (known as the 4B-U-pond) and the region <100 m downstream (Table 2). The 4B-U-pond has a surface area of 2234 m$^2$ and an average depth of 1.1 m. It is located between forested hillslopes, and fed by a forest-covered, cobble-bedded stream. Aerial imagery suggests that the dam at 4B-U pond is between 5-10 years old (Google Earth, 2005; Google Earth 2010). Relative to the 2B-M-pond dam, the 4B-U-pond dam is less soil-packed and has more visible branches, resulting in greater permeability (Figure 4).

Figure 4. The 4B-U-pond dam (left) and pond (right); the photographs were taken on May 27$^{th}$ and May 21$^{st}$ 2019, respectively.
Table 2. Monitoring locations at Site 4B listed by longitudinal distance downstream from the 4B-U-pond monitoring location. The locations in the box are the main study locations.

<table>
<thead>
<tr>
<th>Location</th>
<th>Distance downstream (m)</th>
<th>Depth to bedrock (m)*</th>
</tr>
</thead>
<tbody>
<tr>
<td>4B-U-US-0</td>
<td>-656</td>
<td>1</td>
</tr>
<tr>
<td>4B-U-US-1</td>
<td>-86</td>
<td>4</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>0</td>
<td>3</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>29</td>
<td>2</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>78</td>
<td>5</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>91</td>
<td>5</td>
</tr>
<tr>
<td>4B-L-DS-2</td>
<td>261</td>
<td>4</td>
</tr>
<tr>
<td>4B-L-DS-3</td>
<td>938</td>
<td>0</td>
</tr>
<tr>
<td>4B-L-DS-4</td>
<td>1021</td>
<td>0</td>
</tr>
<tr>
<td>4B-L-DS-5</td>
<td>1065</td>
<td>0</td>
</tr>
</tbody>
</table>

*Jirsa, 2016

A small channel flows from the base of the 4B-U-pond dam through a region of beaver meadow. The first downstream location (4B-U-DS-1) was in this segment, which is shallow and largely unshaded. The stream continues for another 30 m before entering a second beaver pond, 4B-L-pond. The 4B-L-pond dam is older than the upstream pond; aerial imagery suggests that it is close to seventeen years old (Google Earth, 2003). During the previous summer, a section of the 4B-L-pond dam washed out and was then repaired. Consequently, this pond was smaller (surface area = 395 m³) and had higher through-flow than the other ponds (Figure 5).

Figure 5. The 4B-L-pond dam (left) and pond (right); the photographs were taken on May 21st and July 11th 2019, respectively.
The outflowing stream, where monitoring location 4B-L-DS-1 was placed, was approximately as deep as the 4B-L-pond and had a very low velocity during the study period. Approximately 20 m beyond 4B-L-DS-1, the stream channel narrows and becomes deeply incised. At distances greater than 100 m downstream of 4B-U-pond, the forest cover is once again intact. At distances > 950 m downstream (elevation = 239 m, slope = 5%), bedrock becomes prominent in the streambed.

IV. DATA COLLECTION METHODS

1. Monitoring equipment installation

Monitoring equipment was installed at multiple longitudinal locations along the Site 2B and 4B tributaries. The first phase of sensor placement, which took place between late May and early June, 2019, was intended to provide a reach-scale overview of the two sites (Figure 6).

Figure 6. Timeline of equipment placement and monitoring activities at Site 2B and Site 4B during 2019.

To this end, temperature sensors (HOBO Water Temp Pro v2, model U22-001, Onset Corp., Borne, MA) were placed in the ponds, upstream of the ponds, and >100 m downstream from the ponds (Table 3). One temperature sensor was affixed to a shaded tree branch at each site to record air temperature, and additional pressure transducers (HOBO 30-Foot Depth Water Level Logger, model U20-01-001, Onset Corp., Borne, MA) were installed at the pond beds and on a tree branch to record barometric pressure.
A second round of equipment installation took place in mid-July (Figure 6). The purpose of this round was to allow for more intensive monitoring of main study locations, which included the 2B-M-pond, 4B-U-pond, and stream segments <100 m downstream of both impoundments. Beginning on July 16th, pressure transducers were placed at the bed of each location to record water temperature and depth (Table 4). Then, vertical arrays of temperature sensors (Thermochron iButton, model DS1921H F5#, OnSolution Pty Ltd, NSW, Australia) were installed in the shallow subsurface at each location. The vertical arrays – consisting of three temperature sensors affixed to a post and the driven into the streambed – were set to record temperature every fifteen minutes at a depth of 5, 15 and 25 cm below the bed. Shallow wells were also installed in the streambed at each of these locations for later use in tracer testing. The wells were constructed from schedule-40 poly vinyl chloride (PVC) pipe with an inside diameter of 5.08 cm. A length of 10 cm at the base of each well was slotted and screened, and the wells were installed so that this permeable section was 15-25 cm below the streambed or pond bed.

### Table 3. Monitoring equipment installed during phase one at Site 2B and Site 4B.

<table>
<thead>
<tr>
<th>Location</th>
<th>Placement</th>
<th>Equipment</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-US-1</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>2B-M-pond</td>
<td>Pond surface</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>2B-M-pond</td>
<td>Pond bed</td>
<td>Water temp and depth sensor</td>
</tr>
<tr>
<td>Near 2B-M-pond</td>
<td>Tree branch</td>
<td>Air temp sensor</td>
</tr>
<tr>
<td>2B-M-DS-4</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>2B-M-DS-5</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>2B-M-DS-6</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>2B-M-DS-7</td>
<td>Streambed</td>
<td>Water temp and depth sensor</td>
</tr>
<tr>
<td>4B-U-US-0</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>4B-U-US-1</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>Near 4B-U-pond</td>
<td>Tree branch</td>
<td>Air temp sensor</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>Pond bed</td>
<td>Water temp and depth sensor</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>Pond surface</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>4B-L-DS-2</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>4B-L-DS-3</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>4B-L-DS-4</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
<tr>
<td>4B-L-DS-5</td>
<td>Streambed</td>
<td>Water temp sensor</td>
</tr>
</tbody>
</table>
Finally, during this period, two weather stations (HOBO Weather Station 2-Meter Tripod Kit, model # M-TPB-KIT, Onset Corp., Borne, MA) were installed at Site 2B (Figure 7). The first weather station (2B-WS-1) was placed on the stream bank 11 m downstream of the main dam near location 2B-DS-1. The tripod was placed in the beaver forage zone, an area defined by its lack of tree cover. Solar radiation (Silicon Pyranometer Smart Sensor, model S-LIB-M003, Onset Corp., Borne, MA), air temperature and relative humidity (12-bit Temperature/Relative Humidity Smart Sensor, model S-THB-M002, Onset Corp., Borne, MA), and wind speed (Wind Speed
Smart Sensor, model S-WSB-M003, Onset Corp., Borne, MA) were recorded at the tripod weather station at a fifteen-minute interval. A second tripod weather station (2B-WS-2) was placed on a stream bank 37 m downstream of the main dam, near 2B-DS-3. This tripod was located under tree canopy that showed no signs of beaver forage. 2B-WS-2 recorded only solar radiation because air temperature, relative humidity, and wind speed were assumed to be similar at both tripod weather stations. Relative humidity, wind speed, and solar radiation from 2B-WS-1 were also used as a proxy for weather conditions at Site 4B, which has no tree cover downstream of the impoundment.

Figure 7. Weather station 2B-WS-1 (left), weather station 2B-WS-2 (middle), and an array of temperature sensors (right).

2. Site soil characterization

At the main study locations, substrate samples were collected from the streambed to assess soil texture using soil particle size analysis. Dry soil samples were mixed in Calgon solution and a hydrometer was used to measure the solution density relative to the specific density of water (Gee and Bauder, 1986). The hydrometer results were temperature corrected and a calculation was applied to determine the clay, silt, and sand component by percent of total mass. The ratio of clay, silt, and sand was then used to classify each sample by soil texture.

A separate set of soil samples were collected at Site 2B and 4B for bulk density analysis. For these samples, an AMS bulk density corer was used to ensure that the sediment was not disturbed during sample collection. Samples were weighed and oven dried; the wet sample mass $M_{w}$ (kg),
dry sample mass $M_s$ (kg), and known volume of the collection tin $V$ ($9.059 \times 10^{-5}$ m$^3$) were used to characterize the bulk density and other soil properties. Bulk density $\rho_b$ (kg/m$^3$) was calculated as

$$\rho_b = \frac{M_s}{V} \quad \text{Eq. 2}$$

Porosity $\phi$ was then calculated from bulk density and particle density $\rho_s$ (kg/m$^3$), which is assumed to be 2650 kg/m$^3$ based on the density of silica (SiO$_2$).

$$\phi = 1 - \frac{\rho_b}{\rho_s} \quad \text{Eq. 3}$$

Gravimetric water content $\theta_m$ (kg/kg) was determined from wet soil mass and dry soil mass

$$\theta_m = \frac{M_w - M_s}{M_s} \quad \text{Eq. 4}$$

Volumetric water content $\theta_w$ (m$^3$/m$^3$) was found from gravimetric water content, bulk density, and water density $\rho_w$ (kg/m$^3$)

$$\theta_w = \left(\frac{M_w - M_s}{\rho_w}\right) \times \frac{\rho_b}{\rho_w} \quad \text{Eq. 5}$$

Thermal conductivity $\lambda$ (W/m $^\circ$C) was then calculated using Kersten’s empirical model (Kersten, 1949; Rerak, 2017). In this formulation, which has been applied successfully to various soil types, volumetric water content and bulk density were used to derive thermal conductivity.

$$\lambda = (0.13 \log(\theta_w \times 100) - 0.0288) \times 10^{0.000624 \rho_b} \quad \text{Eq. 6}$$

Heat capacity of the soil matrix $C_s$ (J/m$^3$ $^\circ$C) was calculated from the average heat capacity of soil solids $C_{sav}$ (837 J/ kg $^\circ$C) as well as the volumetric heat capacity of water $C_v$ ($4.19 \times 10^6$ J/m$^3$ $^\circ$C) and volumetric water content $\theta_w$ (m$^3$/m$^3$).

$$C_s = (\rho_b \times C_{sav}) + (C_v \times \theta_w) \quad \text{Eq. 7}$$
Thermal diffusivity $\kappa$ (m$^2$/s) was calculated from thermal conductivity and the heat capacity of the soil matrix

$$\kappa = \frac{\lambda}{C_s}$$

Eq. 8

These subsurface soil characteristics, including thermal diffusivity, thermal conductivity, and porosity, were later used to calculate the rate of seepage between the streambed and the stream channel.

3. Stream data

Temperature

In-stream temperature sensors were used to gain a broad understanding of how water temperature may be altered in and around beaver impoundments. Average daily water temperature, temperature difference between upstream and downstream locations, and temperature variance were calculated for each monitoring location for the duration of the July 26th – August 12th study period. At the main study locations (<100 m downstream of dams), temperature was also used to calibrate longitudinal models of stream temperature.

Pond bed or streambed temperature was collected at a fifteen-minute interval at all monitoring locations. The streams were assumed to be well-mixed, so temperature data collected at the streambed was used to represent the average stream temperature of that location. At the pond locations, however, thermal stratification is thought to occur during the summer period. To account for this potential thermal gradient, one temperature sensor was placed just below the water surface at 2B-M-pond and the pond surface at 4B-L-pond. At 4B-U-pond, where surface temperature was not recorded, surface temperature was estimated from the temperature recorded at 4B-L-pond.

Velocity and Discharge

Stream velocity, stream depth, and cross-sectional area were collected at Site 2B and Site 4B stream gauges during the 2018-2019 field season. At Site 2B, the stream gauge is located at 2B-M-DS-7, 340 meters downstream from 2B-M-pond. At Site 4B, the stream gauge is located at 4B-L-DS-4, 1021 meters downstream from 4B-U-pond. These data were used to calculate stream discharge via rating curves for the 2B and 4B streams at each of the gauges.
At both sites, the outflowing stream discharge from the base of the dam was similar to the discharge measured further downstream at the stream gauges. Cross-sectional data were collected at 2B-M-DS-1 and 4B-U-DS-1 so that the stream gauge rating curves could be scaled appropriately. At 2B-M-DS-1 and 4B-U-DS-1 (as well as at the other study locations), cross-sectional area was estimated by multiplying the depth by the wetted width. Then, discharge rates were calculated at 2B-M-DS-1 and 4B-U-DS-1 by multiplying velocity (measured using a stream current meter) by the stream area. These discharge rates were compared to discharge concurrently calculated from the downstream gauge rating curves. Then, the stream gauge rating curves were scaled down in proportion to the channel dimensions of 2B-M-DS-1 and 4B-U-DS-1 and fitted to discharge rates measured at these locations.

The rating curve for 2B-M-DS-1 was applied as far downstream as the 2B-M-DS-3 location (63 m downstream of the 2B-M-pond dam). Likewise, the rating curve developed for 4B-U-DS-1 was applied as far downstream as 4B-L-DS-1 (91 m downstream of the 4B-U-pond dam). Stream velocity was also calculated at each of these locations by dividing the discharge rate by the cross-sectional area.

4. Tracer testing

Tracer testing was completed during the low-flow period in late July 2019 to characterize the hyporheic zone of the two beaver-impounded stream sites. At each site, tracer dye Rhodamine WT was injected into a shallow (30 cm deep) well located in the pond substrate immediately upstream of the beaver dam. Then, dye concentration was measured at three successive downstream locations. Depending on equipment availability, measurements at each location either took place in situ or were periodically collected from within shallow wells and the surface water adjacent to each well.

Tracer testing layout and conditions

At Site 4B, dye was injected into a shallow well at the 4B-U-pond and three downstream locations were subsequently monitored and sampled for dye concentration. The first sampling well location (4B-U-DS-1) was in a relatively shallow and narrow stream section 29 m downstream from the 4B-U-pond dam. The second well location (4B-L-pond) was in a small impounded pond about 78 m downstream of 4B-U-pond, which was approximately 0.5 meters
deep at the time of the test. Finally, the third well location (4B-L-DS-1) was located 91 meters downstream of 4B-U-pond, and almost immediately downstream of the lower pond’s dam. At the time of testing the third location was mostly stagnant and had a depth of 0.5 meters. Site 4B experienced precipitation (<1 cm) the night before the test, but otherwise had been under dry, low-flow conditions for the week prior to testing.

At Site 2B, dye was injected into a shallow well at the 2B-M-pond. The first downstream well location (2B-M-DS-1) was approximately 11 m downstream, in a shallow and narrow channel that flows from the main dam. The second well location (2B-M-DS-2) was in a narrow but deep stream section about 37 m downstream of the main dam. Finally, the third well location (2B-M-DS-3) was located 65 m downstream of the main dam, in a 0.3 deep stream section. No precipitation events >1 cm were recorded in the week prior to testing.

**Tracer testing equipment**

A variety of equipment and techniques were used to determine Rhodamine WT concentration in the stream channel and shallow subsurface at Site 4B. In the stream at 4B-U-DS-1, a field fluorimeter (Seapoint Rhodamine Fluorometer, Seapoint Sensors, Inc., Exeter, NH) was used to log voltage as a proxy for Rhodamine WT concentration at a ten-second interval. Simultaneously, an autosampler was programmed to collect samples from the 4B-U-DS-1 well at a fifteen-minute interval. At the next downstream well, 4B-L-pond, grab samples were collected from the pond surface every 15-30 minutes and a well pump was used to collect samples from the well at the same time interval. Finally, at the 4B-L-DS-1 location, grab samples were collected at the same 15 to 30-minute interval as 4B-U-DS-1.

During the Site 2B tracer test, the fluorimeter was placed in the stream at 2B-M-DS-1 and similarly used to log voltage at a 10 second interval. Also, an autosampler was programmed to collect grab samples from the 2B-M-DS-1 well every 30 minutes. At the second downstream location (2B-M-DS-2), a well pump was used to collect samples from the well, while grab samples were concurrently collected from the surface. Likewise, at the third downstream location (2B-M-DS-3), a well pump was used to collect samples from the well while grab samples were collected from the surface.
Tracer testing timeline

Tracer testing was conducted at Site 4B between July 24th and 25th, 2019. In the hour prior to tracer testing, grab samples were collected from each of the locations as a baseline. At 12:02 pm, approximately 8 oz of 5% concentrate Rhodamine WT dye was injected into the shallow PVC well at 4B-M-pond. Immediately after the dye injection was completed, the autosampler at 4B-U-DS-1 was set to begin sampling and grab sampling was initiated at a 15-minute interval at three downstream locations. Sampling continued consistently until approximately 12:00 am, at which point samples were collected less frequently. The final samples were collected on July 25th at approximately 10:00 am.

Tracer testing at Site 2B was conducted between 7:00 am on July 27th and 10:00 am on July 28th, 2019. In the hour prior to tracer testing, grab samples were collected from each of the three downstream locations to determine baseline concentration. Approximately 14 oz of 5% concentrate Rhodamine WT dye was then injected into the 2B-M-pond well and sampling commenced at the three downstream locations. Based on preliminary results from the 4B tracer test, which suggested that sampling at a 15-minute interval was gratuitous, grab samples were collected at 2B-M-DS-2 and 2B-M-DS-3 every 20 minutes between 7:00 am and 1:00 pm, and then every half-hour between 1:00 pm and 8:00 pm. Autosampling took place at 2B-M-DS-1 every 20 minutes between 7:00 am and 8:00 pm, and then every 30 minutes until 8am the following day. Final samples were collected from each location at 10:00 am on July 28th.

Laboratory testing of tracer samples

All samples collected during the tracer tests at Site 2B and Site 4B were later analyzed in a laboratory setting. To prevent photochemical decay of the Rhodamine WT, the grab samples were covered and refrigerated until this analysis could be completed. Between August and September 2019, a laboratory fluorimeter (TD-700 Laboratory Fluorometer, Turner Designs, Inc., San Jose, CA) was used to analyze grab samples collected during the 4B and 2B tracer tests. The laboratory fluorimeter was calibrated to a series of standards measured at 0.2, 2.0, and 20 ppb Rhodamine WT, which was the approximate range of the collected samples. Based on this calibration, an equation was developed to help determine the concentration of each sample.
Each sample was filtered using a 40-micron nylon syringe filter. Then, at least 4 mL of sample was injected into a quartz cuvette and placed in the laboratory fluorimeter. Samples above the 20 ppb upper limit were diluted to a specified ratio.

**Hydraulic conductivity values**

Rhodamine WT concentration at each sample location varied over the course of the tracer test. Concentration generally increased over time, reached a peak, and then tailed off for the remainder of the sample period. For each sample location, these concentration curves were analyzed to determine the total travel time $t_t$ from the beginning of the tracer test to the peak of the dye plume. The distance traveled over this time is equivalent to Darcy flux $q$ (m/s), the rate of flow through the porous medium.

$$\frac{\Delta l}{t_t} = \frac{\Delta l}{t_p-t_i} = q$$

Eq. 9

where $t_p$ is the peak of the dye plume (s), $t_i$ is the beginning of the tracer test (s), and $\Delta l$ is the horizontal flow path distance between the upstream and downstream location (m). Differences in elevation and distance between upstream and downstream monitoring locations were used to determine hydraulic gradient $i$ (m/m), where $\Delta h$ is hydraulic head difference (m).

$$i = \frac{\Delta h}{\Delta l}$$

Eq. 10

In stream environments, sediment bedding typically results in hydraulic conductivity and flow primarily in the horizontal direction (Freeze and Cherry, 1979; Prych, 1999). In this context, Darcy’s law was used to calculate average hydraulic conductivity $K$ (m/s) between the injection point and each downstream location. Darcy’s law relates flux, hydraulic gradient, and hydraulic conductivity of the sediments (Hatch et al., 2006; Lautz et al., 2006; Briggs et al., 2012) as

$$q = \frac{Q}{A} = Ki$$

Eq. 11

The seepage velocity $v_s$ (m/s), also known as average linear groundwater velocity, is related as the Darcy flux $q$ divided by the effective porosity $n$.

$$v_s = \frac{q}{n}$$

Eq. 12
In the current study, hydraulic conductivity values are assumed to be saturated, unless otherwise indicated. In addition to characterizing the subsurface matrix, tracer testing may be used to evaluate the relative differences in flow rate between the surface stream and the subsurface.

V. HEAT FLUX MODELING METHODS

Heat fluxes at the stream-streambed boundary and stream-atmosphere boundary exert a strong influence over stream thermal regime (Hannah et al., 2004; Webb and Zhang, 1999). The sum of these fluxes was calculated at 2B-M-pond, 4B-U-pond, and study locations <100 m downstream using a heat energy balance equation

\[ h_t = h_e + h_r + h_{se} + h_{bc} + h_a \]  

Eq. 13

where heat flux \( h_t \) (W/m\(^2\)) is the sum of heat flux from evaporation or condensation \( h_e \) (W/m\(^2\)), heat flux from radiation \( h_r \) (W/m\(^2\)), heat flux from sensible heat transfer \( h_{se} \) (W/m\(^2\)), heat flux due to streambed conduction \( h_{bc} \) (W/m\(^2\)), and heat flux from streambed advection \( h_a \) (W/m\(^2\)) (Figure 8). Each term was calculated from a combination of measured values and empirical or physically-based equations.

Figure 8. Diagram of the stream cross section and heat fluxes from evaporation, radiation, sensible heat transfer, streambed conduction, and streambed advection.
1. Stream surface heat and mass flux

Air temperature, wind speed, relative humidity, and solar radiation data from the two weather stations, as well as atmospheric pressure and water temperature, were used to determine heat and mass flux between the water surface and the atmosphere. Vertical heat and mass flux at this boundary were calculated for the ponds and immediate downstream monitoring locations at each site. Given the assumption of vertically well-mixed streams, the temperature sensors used for stream heat calculations were located at the streambed. Because the ponds were thermally stratified, temperature sensors at the pond surfaces were used to calculate heat and mass fluxes at the pond surfaces.

Radiative heat flux

Radiative heat flux $h_r$ is the largest component of the stream heat budget and the most variable on diel timescales. To calculate this heat flux, net shortwave (solar) radiation $S_n$ (W/m$^2$) was added to net longwave radiation $L_n$ (W/m$^2$). Net solar radiation was determined by multiplying incoming solar radiation $S_r$ (W/m$^2$), which was collected at Site 2B, by the proportion of solar radiation that reaches the water $S_w$ (0-1) and the portion absorbed by the water $S_{ab}$ (0-1). An albedo of 0.08 was determined based on criteria outlined by Dickinson (1983), indicating that approximately 92% of incoming solar radiation to the monitored stream reaches was absorbed.

$$h_r = S_n + L_n \quad \text{Eq. 14}$$

$$S_{ab} = (1 - \text{albedo}) \quad \text{Eq. 15}$$

$$S_n = S_r \times S_{ab} \times S_w \quad \text{Eq. 16}$$

The portion of solar radiation that reached the stream was reduced by shading from streambanks and riparian vegetation. The percent of stream width under shade was calculated as a function of solar altitude, azimuth, streambank height, streambank setback distance, and vegetation height (Deas et al., 2003; Li et al., 2012; Loicq at el., 2018). First, the length of the shadow across the stream parallel to the azimuth of the sun $W_p$ (m) was calculated as

$$W_p = \frac{(B_h + V_h)}{\tan(S_{al})} \quad \text{Eq. 17}$$
Where $B_h$ is streambank height (m), $V_h$ is vegetation height (m), and $S_{al}$ is solar altitude (radians) (Figure 9). At locations where bank height and vegetation height differed between the east and west sides of the stream, these values were set to vary as a function of solar azimuth $S_{az}$ (radians) and stream aspect $S_{as}$ (radians), so that

- if $S_{az} + S_{az} < 180$ degrees, then $(B_h + V_h) = $ east bank height
- if $S_{az} + S_{az} > 180$ degrees, then $(B_h + V_h) = $ west bank height

![Figure 9](image-url)  

Figure 9. Diagram of the stream cross section and variables used to calculate the length of a shadow across the stream parallel to the azimuth of the sun.

The length of the shadow normal to the stream $W_n$ (m) was then calculated using $W_p$, as well as the difference between the stream aspect $S_{as}$ (radians) and the solar azimuth $S_{az}$ (radians) (Figure 10).

\[ W_n = W_p \times \sin(S_{az} - S_{as}) \quad \text{Eq. 18} \]
Finally, the proportion of solar radiation reaching the water was calculated from normal shadow width, stream width, and the bank distance $B_d$ (m) setback from the stream.

$$S_w = \left(\frac{W_n - B_d}{W}\right)$$

Eq. 19

Net longwave radiation represents the difference between incoming longwave radiation $L_i$ and outgoing longwave radiation $L_o$ (Oke, 2002). Incoming and outgoing longwave radiation were calculated from air temperature $T_a$ ($^\circ$C), water temperature, atmospheric emissivity $\epsilon_a$, and surface emissivity $\epsilon_s$.

$$L_n = L_i - L_o$$

Eq. 20

$$L_i = \epsilon_a \sigma_{sb} (T_a + 273.2)^4$$

Eq. 21
\[ L_o = \varepsilon_s \sigma_{SB} (T_w + 273.2)^4 + L_i (1 - \varepsilon_s) \]  \hspace{1cm} \text{Eq. 22}

In equations 21 and 22, the Stefan-Boltzmann constant is \( \sigma_{SB} \) \( (5.67 \times 10^{-8} \text{ W/m}^2 \text{ °C}^4) \) and water surface emissivity is 0.97 (Mohseni and Stefan, 1999). Atmospheric emissivity, which was calculated as described by Brutsaert (1975), is a function of air temperature and vapor pressure. This formulation has applied successfully during dry-season conditions (Silva et al., 2019).

\[ \varepsilon_a = 1.24 \left( \frac{e_a}{T_a} \right)^{1/7} \]  \hspace{1cm} \text{Eq. 23}

Saturated vapor pressure \( e_s \) (hPa) was calculated based on the improved Tetens equation (Alduchov and Eskridge, 1996). Actual vapor pressure \( e_a \) (hPa) was calculated from the product of relative humidity and vapor pressure (Silva et al., 2019).

\[ e_s = 6.108 \times e^{17.27T_a/237.3+T_a} \]  \hspace{1cm} \text{Eq. 24}

\[ e_a = e_s \times \frac{RH}{100} \]  \hspace{1cm} \text{Eq. 25}

**Evaporative heat flux**

A stream’s evaporation rate is dependent upon available stream energy and the ability of vapor to be transported away from the water surface. During evaporation, latent heat is lost at the stream surface and cooling occurs. While there are several methods for calculating evaporation rate, evaporative heat flux at the stream surface \( h_e \) (W/m\(^2\)) is typically calculated as

\[ h_e = E_v L_v \rho_w \]  \hspace{1cm} \text{Eq. 26}

\[ L_v = (2454.9 - (2.366 \times T_a)) \times 1000 \]  \hspace{1cm} \text{Eq. 27}

where \( E_v \) is evaporation rate (m/s) and \( L_v \) is the latent heat of vaporization (J/kg) (Dugdale et al., 2017). In several stream modeling scenarios that estimate evaporation rate, Dalton-type equations based on mass transfer are applied. These equations take the general form

\[ E_v = M (e_w - e_a) \]  \hspace{1cm} \text{Eq. 28}

where \( M \) is a mass transfer coefficient and \( e_w \) is saturated vapor pressure based on water temperature (Łaszewski, 2015; Dugdale et al., 2017). The equation may be adjusted empirically.
for atmospheric stability and wind speed; one common formulation is expressed by Webb and Zhang (1999) as

\[ E_v = (1.9 \times 10^{-3})(0.8 + 0.864U)(e_w - e_a) \]  
\text{Eq. 29}

where \( U \) is windspeed (m/s). When the appropriate meteorological data are available, Penman or combination evaporation models (Penman, 1948; Monteith, 1965; Priestly and Taylor, 1972) have also been used to estimate stream surface evaporation rate (Westhoff et al., 2007). In this study, the method formulated by Webb and Zhang (Eq. 29) was applied to find evaporation rate \( E_v \) at each monitoring location. Then, evaporation rate was used to calculate evaporative heat flux using Eq. 26.

**Sensible heat flux**

The sensible heat flux between the air and stream surface is typically a small part of the stream heat budget (Caissie, 2006). Because sensible transfer contributes heat to the stream in proportion to the temperature difference between the air and water, this flux tends to fluctuate on diel time scales, peaking in the middle of the day during the summer. Sensible heat flux was calculated from evaporative heat flux and the Bowen ratio \( B_r \) (Webb and Zhang, 1999) as

\[ h_{se} = h_e B_r \]  
\text{Eq. 30}

\[ B_r = \frac{0.61}{1000} \times \frac{P_a (T_w - T_a)}{e_w - e_a} \]  
\text{Eq. 31}

Where the Bowen ratio is derived from atmospheric pressure \( P_a \) (hPa), water temperature, air temperature, saturated vapor pressure based on water temperature, and actual vapor pressure.

2. Streambed heat and mass flux

Streambed heat flux was calculated as the sum of streambed conduction and advective heat flux from the hyporheic zone. Streambed heat advection and conduction are both derived from vertical temperature gradients in the streambed along with measured soil characteristics.

**Advective heat flux**

Streambed seepage, the advective movement of water mass between the streambed and stream channel, can be found by measuring the extent to which a stream’s temperature is attenuated as a
function of depth in the shallow subsurface (Stallman, 1965; Constantz, 2008). At each sequentially deeper point in the subsurface, temperature signal over time is both attenuated in amplitude and phase shifted. This fluctuation, along with subsurface soil characteristics including thermal diffusivity $\kappa$ (m$^2$/s), thermal conductivity, and porosity, was used to calculate the rate of seepage between the streambed and the stream channel.

Equations for 1-D advective mass flux in the vertical direction were solved using the MATLAB program VFLUX (Gordon et al., 2012; Irvine et al., 2015). VFLUX was also used to filter the raw temperature data and determine the amplitude of the temperature data through dynamic harmonic regression. Streambed seepage was then determined at depths where temperature data were available.

Streambed characteristics are an important control of seepage rate. The sediment effective thermal diffusivity $\kappa_e$ (m$^2$/s) was calculated from effective thermal conductivity $\lambda_e$ (W/m °C) and the heat capacity of the sediment (Hatch et al., 2006; Lautz et al., 2006, Briggs et al., 2012).

$$\kappa_e = \frac{\lambda_e}{c_s}$$  \hspace{1cm} \text{Eq. 32}

The unitless coefficient $\alpha$ was then calculated, where $P_t$ is period of temperature variations ($P_t = \frac{1}{\text{frequency}}$) and $v$ is velocity of the thermal front derived from the amplitude (m/s) (Hatch et al., 2006; Lautz et al., 2006, Briggs et al., 2012).

$$\alpha = \sqrt{v^4 + \left(\frac{9\pi k_e}{P_t}\right)^2}$$  \hspace{1cm} \text{Eq. 33}

Methods developed by Hatch et al. (2006) use a ratio of temperature signal in the subsurface to temperature signal in the stream as a component of this model. This amplitude ratio $A_r$ was calculated using daily maximum $T_{\text{max}}$ (°C) and daily minimum temperature $T_{\text{min}}$ (°C) of the stream and the streambed (Lautz et al., 2006; Briggs et al., 2012).

$$A_r = \frac{T_{\text{max,bed}} - T_{\text{min,bed}}}{T_{\text{max,strm}} - T_{\text{min,strm}}}$$  \hspace{1cm} \text{Eq. 34}

Finally, streambed seepage $q_v$ (m/s) was calculated where $\Delta z$ (m) is difference in depth between a given pair of temperature sensors.

$$q_v = \frac{c_w}{c_v} \left(\frac{2k_e}{\Delta z}\right) \ln A_r + \sqrt{\frac{\alpha + v^2}{2}}$$  \hspace{1cm} \text{Eq. 35}
Streambed seepage was then used to determine heat inputs to the stream. Advective heat flux is calculated as described by Caissie and Luce (2017) as

$$h_a = q_w \rho_w c_w (T_u - T_l)$$  \hspace{1cm} \text{Eq. 36}$$

where vertical seepage is derived from water density, water specific heat, temperature of the upper sensor $T_u$ (°C), and temperature of the lower sensor $T_l$ (°C).

**Conductive heat flux**

Heat flux from streambed conductance is also controlled by bed material and may be derived from temperature monitoring and analysis of streambed sediments. Soil collected at each monitoring location was analyzed to determine thermal conductivity of the bed sediments (Eq. 6). The amount of heat exchanged in bed conduction processes was then calculated based on the work of Theurer et al. (1984) as

$$h_{bc} = -\lambda \frac{\partial T}{\partial z}$$  \hspace{1cm} \text{Eq. 37}$$

where $\frac{\partial T}{\partial z}$ is the vertical streambed temperature gradient (°C/m) at each monitoring location.

### 3. Longitudinal modeling

Longitudinal changes in stream temperature may be modeled from stream discharge, upstream temperature, and the net exchange of heat over this distance. Several studies have implemented the one-dimensional (1-D) heat transport equation to describe thermal transport in streams (MacDonald et al., 2014; Qiu et al. 2020). One well-established formulation of this equation, known as the MNSTREM model, is used in this study to predict stream temperature at varying distances downstream of beaver dams (Sinokrot and Stefan, 1993). In this finite-difference solution to unsteady heat transport,

$$\frac{\partial (AT_x)}{\partial t} + \frac{\partial (QT_x)}{\partial x} = \frac{Wh_t}{\rho_w c_w}$$  \hspace{1cm} \text{Eq. 38}$$

stream temperature is a function of stream width, cross-sectional area $A$, stream discharge, water density, water specific heat, and the sum of the heat flux sources and sinks. This model is designed for free-flowing, well-mixed stream channels where transport is primarily driven by longitudinal advection (dispersion is neglected). The finite-difference discretization outlined in
Herb et al. (2008), which approximates a Taylor series expansion, was used to solve for the temperature of the control volume. The control volume is defined as the discrete physical space (i.e. the stream monitoring location) that heat and mass flux move through. In this discretization scheme

\[ C_1 T_{i,j+1} = C_2 T_{i-1,j} + C_3 T_{i-1,j+1} + C_4 T_{i,j} + C_5 \]  
Eq. 39

\( C_1 \) through \( C_5 \) are discrete cells over which the model is solved, \( i \) is the cell number index in the streamwise direction, and \( j \) is the time step index. The product of \( C_1 \) and \( T_{i,j+1} \) represents the temperature of the control volume at one step forward in time, multiplied by the cell area through which heat and mass are exchanged. Cell \( C_1 \) is solved as

\[ C_1 = A_{i,j+1} = \frac{2 \Delta t Q_{i,j+1}}{\Delta x} - \frac{2 \Delta t W}{\rho_w c_w} \left( \frac{\partial h_i}{\partial T} \right) \]  
Eq. 40

where \( A_{i,j+1} \) is the stream area at one step forward in time (m\(^2\)), \( \Delta t \) is the difference between time steps (s), \( \Delta x \) is the longitudinal stream distance between the control volume and the upstream cell (m), \( Q_{i,j+1} \) is the discharge one step forward in time (m\(^3\)), \( \frac{\partial h_i}{\partial T} \) is the change in stream heat flux over change in stream temperature (W/m\(^2\) °C), and \( W \) is the stream width (m), \( \rho_w \) is water density (kg/m\(^3\)), and \( c_w \) is water specific heat (J/kg °C). Likewise, the product of \( C_2 \) and \( T_{i-1,j} \) represent the cell area multiplied by the temperature of the cell upstream of the control volume at the current time,

\[ C_2 = A_{i-1,j} \]  
Eq. 41

the product of \( C_3 \) and \( T_{i-1,j+1} \) represent cell area (as a function of mass flux over distance) multiplied by the temperature of the cell located upstream of the control volume at one step forward in time,

\[ C_3 = -A_{i-1,j+1} + \frac{2 \Delta t Q_{i-1,j+1}}{\Delta x} \]  
Eq. 42

the product of \( C_4 \) and \( T_{i,j} \) represent cell area (as a function of derived heat flux and width) multiplied by the temperature of the control volume at the current time,

\[ C_4 = A_{i,j} - \frac{2 \Delta t W}{\rho c_p} \left( \frac{\partial h_i}{\partial T} \right) \]  
Eq. 43
and $C_5$ represents the cell area and temperature as a function of absolute heat flux and width.

$$C_5 = \frac{2\Delta t W}{\rho C_p (h_t)}$$  \hspace{1cm} \text{Eq. 44}$$

This upwind scheme allows for temperature to be modeled forward in time, given measured physical parameters and the control volume temperature set as an initial boundary condition for $t=0$. This model formulation was implemented at stream locations 2B-M-DS-1, 2B-M-DS-2, 4B-U-DS-1, and 4B-L-DS-1. For locations immediately downstream of a pond, the pond bed temperature was used as an upstream temperature condition.

**Model calibration**

Streambed temperature measured between July 26th – August 12th 2019 was used to calibrate the model for each location. Throughout this period, average stream temperatures remained relatively consistent. Precipitation occurred on July 29th (3.9 cm) and August 6th (0.9 cm), resulting in slight discrepancies around these dates. Statistical analyses were used to evaluate model performance. First, root-mean-square-error (RMSE) was used to quantify model goodness-of-fit primarily based on absolute numeric agreement. RMSE was calculated as

$$RMSE = \sqrt{\frac{\sum_{i=1}^{n_o} (T_o - T_p)^2}{n_o}}$$  \hspace{1cm} \text{Eq. 45}$$

where $n_o$ is the number of observations, $T_o$ is the measured streambed temperature (°C), and $T_p$ is the predicted stream temperature (°C). A high RMSE value indicates a large difference between measured and predicted stream temperature, while zero indicates that the model estimate is a perfect match. Model calibration was also assessed for Nash-Sutcliffe efficiency (NSE), which is useful for evaluating similarity of form over time (Nash and Sutcliffe, 1970; Krause et al., 2005).

$$NSE = \frac{\sum_{i=1}^{n_o} (T_o - T_p)^2}{\sum_{i=1}^{n_o} (T_o - \bar{T}_o)^2}$$  \hspace{1cm} \text{Eq. 46}$$

where $T_o$ is the observed stream temperature (°C), and $T_p$ is the predicted stream temperature (°C). $NSE$ represents the difference between observed and predicted values normalized by the variance. The value of $NSE$ may fall between $-\infty$ and 1, where 1 is a perfect fit and a value <1 indicates that the mean is a better fit than the model. In general, $NSE$ values > 0.5 are considered a good fit for rainfall runoff modeling (Chiew and McMahon, 1993) and a similar range of $NSE$
values have been described in daily stream temperature modeling studies (Chikita et al., 2010; Tung et al., 2014). Values of 0.36 – 0.75 are considered “satisfactory” in many hydrologic studies conducted on a daily time step (Nash and Sutcliffe, 1970; Motovilov et al. 1999). However, as noted by Moriasi et al. (2007), less strict criteria should be employed when considering shorter temporal (i.e. hourly) scales.

**Sensitivity analysis**

The calibrated models were used to explore how stream temperature might change in response to alterations of the physical environment. In reality, many potential scenarios – such as dam removal / collapse, vegetation regrowth, or changes to the local climate - would result in interconnected changes to the stream’s hydrology. A hypothetical beaver dam collapse might result in more variable discharge rates, lower rates of hyporheic flux, and a higher proportion of water flowing from the warm pond surface. While it is impossible to fully predict the complexities of these scenarios, modeling allows us to investigate the range of temperature effects that each might induce.

In a series of simulations, certain parameters were increased and decreased individually to characterize stream temperature sensitivity. The parameters examined include discharge rate, radiative heat flux, and hydraulic conductivity. The range of values used for each parameter are intended to reflect a realistic range for the given location.

**VI. RESULTS**

1. **Tracer testing**

Downstream of the dam at 4B-U-DS-1, tracer dye concentration was measured using a fluorimeter that was placed in the center of the stream (Figure 11). Shortly after the test was initiated, the dye concentration peaked at 1.2 volts in response to the small portion of dye that had been injected directly into the pond surface water. After an hour, concentration dropped to nearly zero as this dye plume passed 4B-U-DS-1. Then, between hours four and ten of the tracer test, concentration increased steadily to 2.3 volts as subsurface water began to enter the stream. The curve flattened after this point, reaching a peak of 2.4 volts at hour eighteen. Finally, the dye concentration decreased for another six hours, until the end of the tracer test.
Figure 11. Fluorimeter-measured stream voltage during tracer testing at Site 4B (left) and Site 2B (right). The fluorimeter was in the stream at 4B-U-DS-1 (July 23rd – 24th) and 2B-M-DS-1 (July 27th – 28th).

At Site 2B, a small amount of dye was similarly added to the pond surface at the beginning of the tracer test. However, an error in the injection process resulted in dye dripping from the well cap onto the pond surface for the first several hours of the tracer test. The concentration curve captured by a fluorimeter at 2B-DS-M-1 reflects the seepage of this dye into the stream channel over the first eight hours of the tracer test. Then, at hour twelve, dye from the subsurface began to enter the stream. The voltage recorded by the fluorimeter quickly climbed, and then leveled off between hours sixteen and twenty-four. The plume tailed off after this point, as illustrated by the large decrease in recorded concentration that lasted until the end of the twenty-eight hour tracer test (Figure 11).

Many of the samples collected from the stream and subsurface were, unfortunately, unusable due to lab processing difficulties. In other cases, the samples were not collected in a way that captured the extent of the tracer breakthrough curve. For locations sampled >30 m downstream, the tracer test duration was either too short or the plume was too diffuse to be definitively interpreted. However, at locations 2B-M-DS-1 and 4B-U-DS-1, which were both <30 m downstream, the timeline of the breakthrough curve was constrained (Table 5). Stream grab sample data suggest that location 2B-M-DS-1 started at a low baseline concentration (approximately zero), increased between hours four and twelve, and returned to baseline concentration by hour twenty-six. Although this does not fully illustrate the breakthrough curve, it suggests that the peak occurred before twenty-six hours. The subsurface-stream samples collected at 4B-U-DS-1 were less clear, but similarly suggest that dye concentration tailed off before the end of the tracer test, at twenty-four hours.
Table 5. Potential hydraulic conductivity $K$ (m/s) values calculated from the range of subsurface flux $q$ (m/s) values observed during tracer tests at Site 2B and Site 4B.

<table>
<thead>
<tr>
<th>Location</th>
<th>Method</th>
<th>Travel time (h)</th>
<th>Min $q$ (m/s)</th>
<th>Max $q$ (m/s)</th>
<th>Min $K$ (m/s)</th>
<th>Max $K$ (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-DS-1</td>
<td>fluorimeter</td>
<td>12 - 27</td>
<td>$1.1 \times 10^{-4}$</td>
<td>$2.6 \times 10^{-4}$</td>
<td>$4.7 \times 10^{-4}$</td>
<td>$1.1 \times 10^{-3}$</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>autosampler</td>
<td>17 - 28</td>
<td>$1.1 \times 10^{-4}$</td>
<td>$1.8 \times 10^{-4}$</td>
<td>$4.6 \times 10^{-4}$</td>
<td>$7.5 \times 10^{-4}$</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>fluorimeter</td>
<td>14 - 24</td>
<td>$2.8 \times 10^{-4}$</td>
<td>$4.8 \times 10^{-4}$</td>
<td>$1.9 \times 10^{-3}$</td>
<td>$3.2 \times 10^{-3}$</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>autosampler</td>
<td>12 - 24</td>
<td>$2.8 \times 10^{-4}$</td>
<td>$5.6 \times 10^{-4}$</td>
<td>$1.9 \times 10^{-3}$</td>
<td>$3.7 \times 10^{-3}$</td>
</tr>
</tbody>
</table>

The timelines estimated from autosampler and fluorimeter results were then used to constrain estimates of hydraulic conductivity. At Site 2B, travel times and measured site gradient suggest that hydraulic conductivity is between $4.55 \times 10^{-4}$ and $1.06 \times 10^{-3}$ m/s. At Site 4B, travel times and measured site gradient suggest that hydraulic conductivity is between $1.85 \times 10^{-3}$ and $3.7 \times 10^{-3}$. Given the uncertainty regarding the beginning of the breakthrough curve, the minimum hydraulic conductivity value is likely the most accurate. However, even the lowest estimates from the tracer test are quite high, especially at Site 4B.

2. Soil characterization

Soil samples were collected at locations <100 m downstream of the beaver ponds at Site 2B and Site 4B and analyzed to determine soil texture (Table 6). These estimates of soil texture were then used to estimate hydraulic conductivity and porosity based on empirical relationships (Rawls and Brakensiek, 1982; van Genuchten et al., 1991). The samples were also used to calculate soil properties including porosity, thermal conductivity, and thermal diffusivity (Table 7). At Site 2B, sandy loam was the dominant soil texture in the streambed. Sandy loam soils typically have a porosity of 0.435 and a hydraulic conductivity of $3.47 \times 10^{-3}$ m/s. The estimated porosity was generally in agreement with calculated porosity.
Table 6. Soil particle size analysis was used to determine soil texture; estimates of soil porosity \( n \) and hydraulic conductivity \( K \) (m/s) were based on standard values for each soil texture.

<table>
<thead>
<tr>
<th>Location</th>
<th>Soil texture</th>
<th>( n )</th>
<th>( K ) (m/s)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-DS-1</td>
<td>sandy loam</td>
<td>0.435</td>
<td>3.47 \times 10^{-5}</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>sandy loam</td>
<td>0.435</td>
<td>3.47 \times 10^{-5}</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>sandy loam</td>
<td>0.435</td>
<td>3.47 \times 10^{-5}</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>sandy clay loam</td>
<td>0.42</td>
<td>6.30 \times 10^{-5}</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>sandy loam</td>
<td>0.435</td>
<td>3.47 \times 10^{-5}</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>sandy loam</td>
<td>0.435</td>
<td>3.47 \times 10^{-5}</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>sandy clay loam</td>
<td>0.42</td>
<td>6.30 \times 10^{-6}</td>
</tr>
</tbody>
</table>

Sandy loam soil was also the dominant texture at Site 4B, although there was slightly more variation between measured locations at this site. The 4B-U-pond was found to have a sandy clay loam texture, while the downstream channel at 4B-U-DS-1 was more sandy loam. Also, the 4B-L-pond was sandy loam, but its downstream channel 4B-L-DS-1 was sandy clay loam. Consequently, the estimated porosity and the estimated hydraulic conductivity were slightly lower for the 4B-U-pond and 4B-L-DS-1. At Site 4B, as with Site 2B, calculated porosity was in a similar range as estimated porosity.

Table 7. Soil samples at Site 2B and Site 4B were collected for bulk density analysis. Soil wet mass, soil dry mass, and sample tin volume were used to calculate bulk density, thermal conductivity, soil heat capacity, thermal diffusivity, and porosity.

<table>
<thead>
<tr>
<th>Location</th>
<th>( \rho_b ) (g/cm(^3))</th>
<th>( \lambda ) (W/m °C)</th>
<th>( C_v ) (J/m(^3) °C)</th>
<th>( \kappa ) (m(^2)/s)</th>
<th>( n )</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-pond</td>
<td>1.51</td>
<td>1.51</td>
<td>2.76 \times 10^6</td>
<td>5.48 \times 10^{-7}</td>
<td>0.43</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>1.12</td>
<td>0.88</td>
<td>2.54 \times 10^6</td>
<td>3.47 \times 10^{-7}</td>
<td>0.58</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>1.55</td>
<td>1.63</td>
<td>2.87 \times 10^6</td>
<td>5.70 \times 10^{-7}</td>
<td>0.41</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>1.56</td>
<td>1.32</td>
<td>2.15 \times 10^6</td>
<td>6.13 \times 10^{-7}</td>
<td>0.41</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>1.38</td>
<td>1.13</td>
<td>2.23 \times 10^6</td>
<td>5.04 \times 10^{-7}</td>
<td>0.48</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>1.37</td>
<td>1.26</td>
<td>2.75 \times 10^6</td>
<td>4.59 \times 10^{-7}</td>
<td>0.48</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>1.22</td>
<td>1.12</td>
<td>3.23 \times 10^6</td>
<td>3.48 \times 10^{-7}</td>
<td>0.54</td>
</tr>
</tbody>
</table>
3. Stream and air temperature

Stream and air temperature were collected at both Site 2B and Site 4B. Although data were collected between June and October at many locations, the period between July 26th and August 12th is primarily used herein to represent low-flow summer conditions. This period is used for analysis and model calibration unless otherwise stated. One exception is location 2B-M-DS-3, which experienced equipment malfunctions and only had reliable data from July 17th – July 23rd.

Diel temperature cycles

The highest summer air temperature, 37.8 °C, was recorded on July 12th at Site 4B. Air temperature fluctuated daily but remained within a consistent range at both sites from early July to mid-August during the study period (Figure 12).

![Graph of air temperature](image)

Figure 12. Air temperature at Site 2B and Site 4B between July 26th – August 12th. Each air temperature sensor was covered in plastic shielding and affixed to a tree branch.

At both sites, daily average air temperature between July 26th – August 12th was greater than daily average stream temperature at all measured locations. Mean hourly average air temperature during this period ranged between 14 - 26.3 °C at Site 2B, peaking at 1:00 pm (Figure 13). At Site 4B, hourly average air temperature ranged from 13.1 – 26.8 °C and peaked at 2:00 pm.
Between July 26th – August 12th, stream temperature had a smaller range of diel variation than air temperature (Figure 13). Stream temperature generally peaked in the early afternoon, corresponding to measured peaks in air temperature. At night, stream temperature decreased slowly, reaching a minimum at approximately 5:00 am. At Site 2B, the diel temperature signal was stronger at stream locations >100 m downstream from 2B-M-pond. At 110, 300, 327, and 340 meters downstream, temperature variance $s^2$ was between 2.09 – 3.07. By contrast, the temperature signals <100 m downstream were more attenuated ($s^2 = 0.32 - 0.59$). At Site 4B, locations >100 m downstream showed a moderate degree of variance (1.66 - 5.33) as did stream locations <100 m downstream (1.47 – 3.55).

The ponds at Site 2B and 4B showed a high degree of diel variation at the surface, but considerably less variation at the bed (Figure 14). For example, while the 2B-M-pond showed a strong diel temperature signal at the surface ($s^2 = 7.93$), the pond bottom temperature remained
stable ($s^2 = 0.05$). Likewise, the 4B-U-pond bed was very stable ($s^2 = 0.63$). At the lower 4B-L-pond, the surface and bed had similar average temperatures but a notable difference in diel fluctuation, the surface variance of 8.75 was the highest recorded at any location.

![Graph of stream temperature](image)

Figure 14. Pond bed and pond surface temperature at Site 4B (top) and Site 2B (bottom) between July 26th – August 12th, 2019.

**Longitudinal stream temperature**

One objective of this study is to understand the extent to which beaver dam-altered stream temperature persists longitudinally downstream. The stream temperature change from upstream to downstream was examined by calculating the minimum, mean, and maximum temperature at each monitoring location, as well as the variance (Figures 15 and 16). These data reveal that the effects of beaver dams in the KRW are highly inconsistent, with some downstream reaches warming and others cooling. The longitudinal distance downstream at which temperature approached the range of upstream temperature was also noted.

At Site 2B, longitudinal temperatures were analyzed during July 17th – July 22nd, 2019. This range was used due to sensor failures at location 2B-U-DS-3 during the July 26th – August 12th period. Although water temperatures were about 0.2 °C warmer on average during the earlier time period, the relationship between sensors was consistent with the later time. Downstream of the 2B-M-pond, stream temperature approached the upstream average temperature by approximately 327 m (-1 °C difference). At Site 4B, the average temperature downstream of the 4B-U-pond closely matched the furthest upstream temperature (-0.3 °C difference) by 261 m.
Figure 15. Average, maximum, and minimum streambed or pond bed temperature as a function of distance downstream. Values were observed at Site 2B between July 17th – July 22nd, 2019.

The average streambed temperature measured at 2B-M-US-1, located 187 m upstream of the pond monitoring location, was 19.4 °C (Figure 15). Relative to this upstream temperature, the 2B-M-pond was cooler at the bed (18.4 °C) but considerably warmer at the surface (24.3 °C). This 6.1 °C thermal gradient was observed over a depth of approximately 1.2 m during the study period. Downstream of the dam the stream remained quite cool. Location 2B-M-DS-1 had an average temperature of 18.4 °C and 2B-M-DS-2 had an average temperature of 16.8 °C. The 2B-M-DS-3 location was 16.9 °C on average. Beyond this point, stream temperature increased by approximately 1°C and stayed consistent (17.8 – 18.4 °C) for the next 110-340 m downstream.

Longitudinal temperature dynamics were distinctly different at Site 4B (Figure 16). Most notably, the first location downstream of the 4B-U-pond was warmer, on average, than the pond bed. Furthermore, this 2.4 °C increase was the only instance of warming measured immediately downstream of a beaver impoundment. The 4B-L-pond exhibited a slight difference in average temperature between the surface (22.1 °C) and pond bed (21 °C). Finally, the temperature immediately downstream of 4B-L-pond was cooler than the pond bed, with an average daily temperature of 20.2 °C. Beyond this point stream temperature decreased slightly, approaching the thermal range of upstream temperature at 261 m downstream.
Figure 16. Average, maximum, and minimum streambed or pond bed temperature as a function of distance downstream. Values were observed at Site 4B between July 26th – August 12th, 2019.

4. Heat flux to the stream: atmospheric processes

Radiative heat flux

Total heat flux to the stream surface was calculated as the sum of net radiative heat flux, evaporative heat flux, and sensible heat flux. At both sites, radiation was the largest heat flux component as well as the most variable on diel time scales. During the study period, incoming solar radiation measured at 2B-WS-1 typically peaked at 11:30 am (Figure 17). Although higher rates of radiation were often recorded in the afternoon, more variable weather patterns during this time of day resulted in lower average radiation.

Average incoming solar radiation measured in direct sunlight at 2B-WS-1 was 179.8 W/m², while average incoming solar radiation measured under the forest canopy at 2B-WS-2 was 40.1 W/m². Direct incoming solar radiation from 2B-W-1 was used as a baseline for every location except for 2B-DS-3, which was under tree canopy and therefore modeled from solar radiation at 2B-WS-2. Net radiation was thus calculated from shortwave and longwave radiation at each study location (Figure 17).
Figure 17. Average hourly incoming solar radiation measured at 2B-WS-1 and 2B-WS-2 (left), and average net radiative heat flux at each monitoring location (right) during July 26th – August 12th, 2019.

The 2B-M-pond and 4B-U-pond both experienced high levels of net radiation, as did the first downstream location from each of these ponds. At 2B-M-DS-2 and 4B-L-DS-1, where riparian vegetation was greater, net radiation was reduced. The 2B-M-DS-3 location, which was under forest canopy, had the lowest average net radiation.

**Evaporation and sensible heat flux**

Latent heat flux due to evaporation and condensation varied greatly on diel time scales. Evaporation rates were greatest in the early afternoon when relative humidity was low and solar radiation and windspeed were at their peak. At night, when solar radiation decreased and relative humidity increased, condensation occurred at the water surface. Site 4B and the 2B-M-pond had high rates of evaporation during the day, while the channel downstream of the 2B-M-pond had lower rates of net evaporation. Evaporation was associated with heat loss at the stream surface, while condensation was a source of heat (Table 8).
Table 8. Average sensible ($h_{se}$) and evaporative ($h_e$) heat flux at each monitoring location during July 26th – August 12th, 2019. Positive values indicate a heat source to the water; negative values indicate a heat sink.

<table>
<thead>
<tr>
<th>Location</th>
<th>$h_{se}$ (W/m$^2$)</th>
<th>$h_e$ (W/m$^2$)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-pond</td>
<td>-11.3</td>
<td>-53.9</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>6.4</td>
<td>-6.3</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>12.6</td>
<td>6.2</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>11.8</td>
<td>2.2</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>-6.6</td>
<td>-40.2</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>-3.8</td>
<td>-31.8</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>-6.6</td>
<td>-40.17</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>-1.1</td>
<td>-24.5</td>
</tr>
</tbody>
</table>

Sensible heat flux, driven by conduction and convection between the air and water, was a comparatively small – although still significant - part of the heat budget. Sensible heat flux similarly transferred heat to the streams at Site 2B and resulted in net heat loss at all other locations. Rates of sensible heat flux fluctuated on diel time scales, with the highest positive rates occurring during the day. However, as with evaporation, this flux was sometimes limited by vapor pressure or relative humidity conditions. The highest average rates of sensible heat flux, 12.6 W/m$^2$ and 11.8 W/m$^2$, were measured at 2B-M-DS-2 and 2B-M-DS-3, respectively.

5. Heat flux to the stream: streambed processes

Heat fluxes at the streambed varied primarily as a function of water mass upwelling or downwelling across the streambed boundary. Temperature was measured at 0, 5, 15, 25, and 30 cm deep in the subsurface, and advective vertical flux rate was calculated for the depth range of each sensor pair (Figure 18). Seepage was generally found to flow downward (recharge) upstream of dams and flow upward (discharge) at locations downstream of dams.
Figure 18. Temperature measured at 5 and 15 cm below the streambed (top) and calculated vertical flux from the hyporheic zone in this range (bottom) at 4B-U-DS-1. Negative values (bottom) indicate upwelling seepage.

Effective upwelling or downwelling flux to the stream is most accurately represented by measurements directly below the streambed. Advective seepage flux magnitude is typically greatest in this range and decreases with depth. However, fluxes measured in this range also have the greatest potential for error due to streambed scour or disturbance. Furthermore, certain depths measured in this study (particularly at the very shallow or deep range) were discarded due to poor fitting between the data and model results. In the later development of 1-D longitudinal stream temperature models, rates of advective heat flux between the hyporheic zone and the stream or pond were based on data collected at average depths ≤ 15 cm.

For some measurements ≥ 20 cm, advective seepage was strongly attenuated or even showed signs of flow in the opposite direction (Table 9). These fluxes reversals may be due to subsurface heterogeneity or potential recharge to groundwater. While deep hyporheic fluxes are relevant to long-term groundwater storage and deep flow paths, they have a minimal impact on stream temperature at the scale of the current study.
Table 9. Average pond bed and streambed mass advection for multiple depths at Site 2B and Site 4B study locations between July 26th – August 12th, 2019. Negative values indicate upwelling seepage; positive values indicate downwelling.

At Site 2B, calculated seepage rates suggest the occurrence of downwelling in the impoundment and upwelling at locations downstream. An average advective downwelling flux of 0.59 m/day was calculated for the 2B-M-pond at a depth of 5-25 cm below the streambed (Figure 19). Downstream at 2B-M-DS-1, average flux upwelling into the stream was 0.73 m/day at a depth of 0-5 cm. Further downstream at 2B-M-DS-2, flux upwelling rate was 0.48 m/day at a depth of 5-25 cm. Although these rates are small, it is important to note that in-stream flows measured during this period were also quite low. Advection was not calculated at 2B-M-DS-3 due to equipment failures.
Figure 19. Pond and streambed mass advection measured near the streambed surface at Site 2B study locations between July 26th – August 12th, 2019. Negative values indicate downwelling seepage; positive values indicate upwelling.

While average advective downwelling in the 4B-U-pond was quite low (0.18 m/day at 0-5 cm), advective upwelling downstream was similar in magnitude to the upwelling at Site 2B. At location 4B-U-DS-1, an upwelling rate of 0.44 m/day was calculated at a depth of 5-25 cm (Table 9, Figure 20). Further downstream at 4B-L-pond, an upwelling rate of 0.45 m/day was calculated at a depth of 5-15 cm. Location 4B-L-pond was the only pond where consistent upwelling was measured. Finally, 4B-L-DS-1 was also found to have upwelling at a depth of 5-15 cm (0.28 m/day).
Figure 20. Pond and streambed mass advection measured near the streambed surface at Site 4B study locations between July 26th – August 12th, 2019. Negative values indicate downwelling seepage; positive values indicate upwelling.

Advective seepage flux rates were used to calculate advective heat transfer to the stream at each location (Table 10). At locations that experienced upwelling, a negative heat flux resulted from the addition of cold water to the stream. At locations that experienced net downwelling – i.e. the 2B-M-pond and the 4B-U-pond – a positive heat flux was observed as cold water left the system. This positive heat flux was very small at 4B-U-pond (2.0 W/m²), and more substantial (32.8 W/m²) at 2B-M-pond. While advective heat flux rates recorded at Site 2B were relatively stable over diel timescales, Site 4B displayed more fluctuation.
Table 10. Average heat fluxes from streambed conduction \( (h_{bc}) \) and advection \( (h_a) \) during July 26\textsuperscript{th} – August 12\textsuperscript{th}, 2019. Positive values indicate a heat source to the stream; negative values indicate a heat sink.

<table>
<thead>
<tr>
<th>Location</th>
<th>( h_{bc} ) (W/m(^2))</th>
<th>( h_a ) (W/m(^2))</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-pond</td>
<td>-13.2</td>
<td>32.8</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>-8.2</td>
<td>-22.1</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>-6.3</td>
<td>-24.4</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>-6.5</td>
<td>2.0</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>-14.9</td>
<td>-32.4</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>-18.8</td>
<td>-38.9</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>-18.5</td>
<td>-10.7</td>
</tr>
</tbody>
</table>

Heat flux from streambed conduction varied as a function of thermal conductivity and water temperature gradient in the shallow subsurface. As with calculations of advective heat flux, temperature gradients in the subsurface resulted in variation depending on the range of depth measurements used. Therefore, to maintain consistency, temperature data collected at 5-15 cm were used for all streambed conduction calculations. At Site 2B, streambed conduction contributed a relatively small negative heat flux throughout the study period. Conductive heat fluxes were greater at Site 4B as well as more variable on diel timescales, particularly in the 4B-L-pond and 4B-L-DS-1.

### 6. Total heat budget

During the study period, the water surface was a net source of heat at all stream monitoring locations (Table 11). Net heat flux at the stream surface ranged from 20.7 W/m\(^2\) (4B-L-DS-1) to 112.1 W/m\(^2\) (2B-M-DS-1). The streambed, by contrast, was typically a heat sink. Average heat losses at the streambed ranged between -29.2 W/m\(^2\) (2B-M-DS-2) and -47.3 W/m\(^2\) (4B-U-DS-1). Each of the ponds experienced a net positive surface heat flux (36.6 – 41.0 W/m\(^2\)), although these rates were generally lower than the stream surface heat flux rates. At the pond beds, heat fluxes were more variable between the impoundments. The 2B-M-pond, which experienced strong advective downwelling, had a net heat gain at the pond bed (19.6 W/m\(^3\)); this was the only location where the bed acted as a net source. The 4B-U-pond had a slightly negative net heat flux at the pond bed (-4.5 W/m\(^3\)), while the 4B-L-pond – the only pond with significant advective upwelling – had a considerably greater degree of cooling at the pond bed (-57.7 W/m\(^3\)).
Table 1. Average net heat flux at the streambed and stream surface, as well as the net total of heat fluxes at these two boundaries, during July 26th – August 12th, 2019.

<table>
<thead>
<tr>
<th>Location</th>
<th>Net streambed flux (W/m²)</th>
<th>Net stream surface flux (W/m²)</th>
<th>Net total flux (W/m²)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-pond</td>
<td>19.6</td>
<td>41.0</td>
<td>61</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>-30.4</td>
<td>112.1</td>
<td>82</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>-30.7</td>
<td>89.7</td>
<td>59</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>-</td>
<td>44.1</td>
<td>-</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>-4.5</td>
<td>37.6</td>
<td>33</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>-47.3</td>
<td>73.8</td>
<td>27</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>-57.7</td>
<td>36.6</td>
<td>-21</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>-29.2</td>
<td>20.7</td>
<td>-9</td>
</tr>
</tbody>
</table>

The net total heat flux, comprised of heat fluxes at both the streambed and stream surface, was positive at most stream locations. Additionally, the positive net heat fluxes at the 2B Site stream locations were generally greater than those at Site 4B. The 2B locations – including the first downstream location 2B-M-DS-1, which was 82 W/m² on average – each had relatively high positive surface heat fluxes, which contributed to the net total flux. The locations at Site 4B that had low or negative net total heat fluxes (i.e. 4B-L-pond and 4B-L-DS-1) had small heat fluxes at the surface and large negative fluxes at the bed.

Variation between the two sites was also apparent in the magnitude of the net heat fluxes from conduction, streambed advection, sensible heat transfer, and evaporative heat flux (Figure 21, Figure 22, Table 12).
Figure 21. Average heat fluxes at Site 2B from radiation, conduction, advection, sensible heat, and evaporative heat. Values are calculated from measurements taken during July 26th – August 12th, 2019.

Each of the stream locations at Site 2B had very low or slightly positive evaporative heat fluxes, while evaporative heat losses were readily apparent at the Site 4B stream locations. The warmer 4B stream had both higher evaporation rates and higher rates of evaporative cooling at the surface. Evaporation accounted for 4% of absolute heat flux at each of the Site 2B stream locations, and 17-24% of absolute heat flux at the Site 4B locations.
Differing rates of sensible heat transfer between Site 2B and Site 4B were also observed. Sensible heat flux at the stream surface is largely driven by air-water temperature gradients; larger gradients observed at Site 2B were associated with positive sensible heat flux from the warm air to the relatively cool stream. At Site 4B, where the stream was warmer, less heat was transferred along this gradient. Sensible heat transfer was a small part of the heat budget at both sites and generally accounted for no more than 5% of total heat flux at a given location, apart from location 2B-M-DS-2 (12%).

Table 12. Heat fluxes as average net total (W/m²) and as percent of absolute average total (%) at each monitoring location during July 26th – August 12th, 2019.

<table>
<thead>
<tr>
<th>Location</th>
<th>$h_r$ (W/m²)</th>
<th>$h_r$ (%)</th>
<th>$h_{se}$ (W/m²)</th>
<th>$h_{se}$ (%)</th>
<th>$h_e$ (W/m²)</th>
<th>$h_e$ (%)</th>
<th>$h_{bc}$ (W/m²)</th>
<th>$h_{bc}$ (%)</th>
<th>$h_a$ (W/m²)</th>
<th>$h_a$ (%)</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-pond</td>
<td>106.2</td>
<td>49%</td>
<td>-11.3</td>
<td>5%</td>
<td>-53.9</td>
<td>25%</td>
<td>-13.2</td>
<td>6%</td>
<td>32.8</td>
<td>15%</td>
</tr>
<tr>
<td>2B-M-DS-1</td>
<td>111.8</td>
<td>72%</td>
<td>6.2</td>
<td>4%</td>
<td>-6.0</td>
<td>4%</td>
<td>-8.2</td>
<td>5%</td>
<td>-22.1</td>
<td>14%</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>71.7</td>
<td>60%</td>
<td>12.8</td>
<td>11%</td>
<td>5.2</td>
<td>4%</td>
<td>-6.3</td>
<td>5%</td>
<td>-24.4</td>
<td>20%</td>
</tr>
<tr>
<td>2B-M-DS-3</td>
<td>31.2</td>
<td>-</td>
<td>11.4</td>
<td>-</td>
<td>1.5</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
<td>-</td>
</tr>
<tr>
<td>4B-U-pond</td>
<td>84.3</td>
<td>60%</td>
<td>-6.4</td>
<td>5%</td>
<td>-40.3</td>
<td>29%</td>
<td>-6.5</td>
<td>5%</td>
<td>2.0</td>
<td>1%</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>109.5</td>
<td>57%</td>
<td>-3.7</td>
<td>2%</td>
<td>-31.9</td>
<td>17%</td>
<td>-14.9</td>
<td>8%</td>
<td>-32.4</td>
<td>17%</td>
</tr>
<tr>
<td>4B-L-pond</td>
<td>83.3</td>
<td>44%</td>
<td>-6.4</td>
<td>3%</td>
<td>-40.3</td>
<td>21%</td>
<td>-18.8</td>
<td>10%</td>
<td>-38.9</td>
<td>21%</td>
</tr>
<tr>
<td>4B-L-DS-1</td>
<td>46.3</td>
<td>46%</td>
<td>-1.0</td>
<td>1%</td>
<td>-24.6</td>
<td>24%</td>
<td>-18.5</td>
<td>18%</td>
<td>-10.7</td>
<td>11%</td>
</tr>
</tbody>
</table>

At Site 2B stream locations, advection was responsible for 14-20% of absolute heat flux. Advection was a similar portion of the total heat fluxes at Site 4B streams (11-17%), and absolute magnitude of advective fluxes were similar between the sites. The 4B-U-pond had the lowest portion of advective heat flux (1%), which represented a heat source to the pond. Streambed conductance was a small but not insignificant component. It was generally responsible for 5-8% of total heat fluxes at a given location, although at Site 4B it was as high as 18% of the absolute heat flux.
7. Model results

At Site 2B, stream temperature was simulated at locations 2B-M-DS-1 and 2B-M-DS-2 (Figure 23, Figure 24). On average, the predicted temperature at 2B-M-DS-1 is 0.01 °C smaller and has a 0.01 °C greater standard deviation than actual temperature. The predicted average maximum daily temperature is 17.99 °C, which is 0.05 °C warmer than observed. At location 2B-M-DS-2, predicted temperature is 0.01 °C greater and has a 0.04 °C smaller standard deviation than the simulation. The maximum daily temperature is 15.51 °C at this location, 0.10 °C cooler than the predicted maximum daily temperature. At both 2B locations, a low-frequency, long-term (weekly) temperature signal is evident in the curvature of the simulation; however, a diel stream temperature signal is not readily apparent. The RMSE values for both Site 2B locations was <0.1 °C and the NSE values were 0.13 and 0.27 for 2B-M-DS-1 and 2B-M-DS-2, respectively (Table 13).

![Graph of temperature data](image)

Figure 23. Simulated and observed stream temperature at 2B-M-DS-1 during July 26th – August 12th, 2019.
At Site 4B, the simulated stream temperature signals are similar in shape, timing, and magnitude to the observed stream temperature signals. At 4B-U-DS-1, the observed temperature is 0.21 °C greater than the predicted temperature on average, with a 0.29 °C smaller standard deviation (Figure 25). The maximum daily temperature is slightly overestimated by the model, by about 0.33 °C on average. The 4B-L-DS-1 predicted and observed temperature signals are likewise quite similar; this location’s observed average temperature is 0.11 °C cooler and has a standard deviation 0.37 °C greater than the predicted stream temperature (Figure 26). On average, the observed maximum daily temperature is 0.08 °C warmer than that of the model at 4B-L-DS-1. At both 4B-U-DS-1 and 4B-L-DS-1, a diel and long-term temperature signal is detectable. The RMSE values for both Site 4B locations was <0.1, and the NSE values were 0.41 and 0.61 for 4B-U-DS-1 and 4B-L-DS-1, respectively (Table 13).
For location 4B-U-DS-1, both the RMSE and NSE values indicate that the model reasonably describes the observed stream temperatures (Chikita et al., 2009). Moreover, 4B-U-DS-1 shows good agreement between the predicted and observed temperature signal across a relatively wide
temperature range (s = 3.5 °C). For 4B-L-DS-1, the RMSE and NSE values similarly within a reasonable range, and the model is a good fit to the stream’s diel temperature signal. By contrast, the two Site 2B models have low (<0.1) RMSE values but are less fitted to the stream diel temperature signal, as indicated by low NSE values. While the observed stream temperature ranges at 2B-M-DS-1 and 2B-M-DS-1 were highly attenuated, the simulated temperature ranges displayed greater diel variation. Furthermore, the stream temperatures at 2B-M-DS-1 and 2B-M-DS-2 appear to be phase shifted relative to upstream temperature.

Table 13. Statistical RMSE and NSE values calculated for each model during the calibration period of July 26th – August 12th, 2019.

<table>
<thead>
<tr>
<th>Location</th>
<th>RMSE</th>
<th>NSE</th>
</tr>
</thead>
<tbody>
<tr>
<td>2B-M-DS-1</td>
<td>0.02</td>
<td>0.13</td>
</tr>
<tr>
<td>2B-M-DS-2</td>
<td>0.02</td>
<td>0.27</td>
</tr>
<tr>
<td>4B-U-DS-1</td>
<td>0.09</td>
<td>0.41</td>
</tr>
<tr>
<td>4B-L-DS-2</td>
<td>0.05</td>
<td>0.61</td>
</tr>
</tbody>
</table>

8. Model sensitivity analysis

Simulations were performed for location 4B-U-DS-1 to understand stream temperature sensitivity to a variety of model parameters, including radiative heat flux, hydraulic conductivity, and discharge rate. This site was selected because its longitudinal model has a good fit with observed stream temperatures. The results of these simulations can be used to gain insight into how parameters impact stream temperature at the other study locations.

At location 4B-U-DS-1, simulations were performed to assess the influence of discharge rate on stream temperature. Temperature was simulated where stream discharge was 60 – 140% of the flow observed during the model calibration period (Figure 27). From this model, it is evident that low discharge rates have the potential to dramatically increase stream temperature at 4B-DS-1. When discharge rate was simulated at 60% of the observed flow, stream temperature increased by 1.4 °C. At 140% of the observed flow, stream temperature was 0.6 °C cooler on average.
Upstream temperature strongly influences stream temperature at a given study location. In simulations of stream temperature at 4B-U-DS-1 in response to 1°C warmer upstream temperature (approximately 11 m upstream), the resulting stream temperature at 4B-U-DS-1 was 1.1 °C warmer on average (Figure 28). Likewise, when upstream temperature was simulated as 1 °C cooler, the resulting temperature at 4B-U-DS-1 was 1.1 °C cooler on average. The relationship between upstream and study location temperature is not one-to-one, due to the dependence of evaporative, radiative, and sensible heat transfer processes on water temperature.

Figure 27. Simulation of stream temperature at 4B-U-DS-1 in response to varying levels of discharge between 60-120% of the observed discharge rate.

Figure 28. Simulation of stream temperature at 4B-U-DS-1 in response to varying upstream temperature up to +/- 2 °C from observed upstream temperature.
Evaporative, radiative, and sensible heat transfer processes are also dependent upon air temperature. In the Lake Superior region, where air temperature is predicted to increase 3 – 4.5 °C by the end of the 21st century (Huff and Thomas, 2014), stream temperatures may be impacted accordingly. When simulated air temperature was increased by 2 °C, the predicted stream temperature at 4B-U-DS-1 increased by 0.3 °C on average (Figure 29). A simulated air temperature increase of 5 °C resulted in 0.6 °C warmer stream temperatures at 4B-U-DS-1.

![Figure 29](image)

Figure 29. Simulation of stream temperature at 4B-U-DS-1 in response to varying air temperature up to +/- 5 °C from observed air temperature.

Vegetative shading at the 4B-U-DS-1 stream location was minimal during the study period. However, vegetation growth in future years could decrease the amount of solar radiation that reaches the stream. Simulations suggest that increasing vegetation height (by 1 m) on both sides of the stream could decrease stream temperature by 0.3 °C (Figure 30). This simulation assumes that the vegetation is set back from the stream by approximately 2 m, as observed during the study period. In simulations of increased vegetation height with decreased set-back distance, stream temperature cooling is greater.
Figure 30. Simulation of stream temperature at 4B-U-DS-1 in response to varying riparian vegetation height between 0-1 m greater than the observed vegetation height.

Stream temperature was also simulated where radiative heat flux to the stream surface was 40-120% of observed values at 4B-U-DS-1 (Figure 31). Decreasing the heat flux from solar radiation had the effect of lowering average stream temperature. When radiation was decreased to 60%, stream temperature decreased by 0.2 °C on average. For radiative fluxes simulated at 40% of observed levels, stream temperature decreased by 0.3 °C. These scenarios suggest how stream temperature may be affected by vegetation regrowth in the beaver forage zone, or by future climate scenarios.

Figure 31. Simulation of stream temperature at 4B-U-DS-1 in response to varying levels of net radiation between 40-120% of observed levels.
Finally, stream temperature was simulated at 4B-U-DS-1 where hydraulic conductivity was 40-400% of the value indicated by soil sample analysis (Figure 32). There is a positive linear relationship between hydraulic conductivity and vertical advective flux, and increasing hydraulic conductivity has the effect of increasing advective heat exchange. Simulations of increased hydraulic conductivity show a stream cooling effect at these upwelling locations. When hydraulic conductivity is decreased to 80% of observed values, average stream temperature increases by 0.1 °C over the simulated distance. At 200% of observed values, average stream temperature decreases by 0.3 °C.

Figure 32. Simulation of stream temperature at 4B-U-DS-1 in response to varying hydraulic conductivity between 40-400% of observed values.

9. Model validation

The 4B-U-DS-1 model was validated by simulating stream temperature for two additional time periods outside of the July 26th – August 12th calibration period. Stream temperature was simulated from July 21st – July 25th, as well as from August 13th – August 17th (Figure 33, Figure 34). The model formulation used in the validation and calibration periods is identical, with one exception; estimates of heat flux due to advective seepage and streambed conduction during the validation periods are based on values from the calibration period, due to limited streambed temperature data.
In the initial validation effort, estimates of heat flux due to advective seepage and streambed conduction are based on average values from the calibration period (Figure 33). For July 21st – July 25th, the fit between the observed temperature and model is better than the fit during the calibration period. This is reflected in the low RMSE (0.04) and high NSE value (0.74) for July 21st – July 25th. The average temperature difference between observed and predicted temperature is comparable (observed temperature is 0.26 °C warmer on average). However, when this formulation is applied during the August 13th – August 17th period, the RMSE value is also 0.04 but the NSE value is lower than during the calibration period (0.31). Stream temperature during the August validation is overestimated by 0.81 °C on average, and maximum daily stream temperature is overestimated by 0.38 °C on average.

In the second validation effort, estimates of advective seepage and streambed conduction are based on hourly average values from the calibration period (Figure 34). In this formulation, the statistical fit of the July 21st – July 25th period is the same as the fit during the initial model validation (RMSE = 0.04, NSE = 0.74), while the fit of the August 13th – August 17th period is improved (RMSE = 0.04, NSE = 0.41). The August validation fit is further improved when simulated heat fluxes due to advective seepage and streambed conduction are increased by 50% (RMSE = 0.04, NSE = 0.52).
Figure 3.4. Simulated and observed temperature during July 21st – July 25th, 2019 (left) and August 13th – August 17th, 2019 (right). In this model, estimates of heat flux due to advective seepage and streambed conduction are based on hourly average values from the July 26th – August 12th calibration period.

VII. DISCUSSION

Beaver impoundments are known to greatly increase geomorphic heterogeneity and water temperature variability at multiple spatial scales (Majerova et al., 2015). This is consistent with findings in the KRW, where measured temperatures during the study period fluctuated up to 12.8 °C within a single geomorphic unit (4B-L-pond), as well as 11.8 °C between units along the same stream reach (2B-pond surface and 2B-M-DS-2). In the current study, beaver impoundments were also associated with regions of extremely low temperature variability. At 2B-M-DS-1, located 37 m downstream of the 2B-M-pond, temperature variance was just 0.6 during the 16-day period. These observations are also in line with previous studies, which attribute such low-fluctuation regions to the buffering effect of dam-induced hyporheic upwelling (Weber et al., 2017). I contend that low-fluctuation locations in the KRW (observed <100 m downstream) were similarly buffered by beaver impoundments. This is further supported by greater stream fluctuations observed >100 m downstream.

Immediately downstream of the beaver ponds, stream temperatures often deviated noticeably from temperatures recorded upstream in free-flowing channels with minimal beaver activity. Despite these impacts, altered stream temperatures did not seem to propagate for great distances downstream. At Site 2B, downstream temperature closely matched upstream temperature (~0.6
°C) by 327 m downstream of the 2B-M-pond. At Site 4B, stream temperature seemed to recover (-0.2 °C) by 261 m downstream of the 4B-U-pond. Alternatively, it is possible that beaver-altered downstream temperatures were not evident due to high variability in the free-flowing downstream channel. In either case, these observations demonstrate the importance of spatial scale when considering the impacts of a beaver impoundment on temperature variation.

Some studies have found that beaver ponds - and chains of beaver ponds, even more so - warm the downstream channel (Margolis et al., 2001). Others have observed warming downstream of certain beaver dams and cooling downstream of others, in relation to site-specific hydrogeologic features (Fuller and Peckarsky, 2011). In observations of beaver ponds in the KRW, downstream warming or cooling trends were similarly site-specific. Downstream of 2B-M-pond and 4B-L-pond, average stream temperature was lower than average pond bed temperature. By contrast, average stream temperature downstream of the 4B-U-pond was significantly warmer than average pond bed temperature. Furthermore, the 2B-M-pond (which produced the coolest measured downstream temperatures) was at the downstream end of an extensive beaver pond complex.

The differing impact of these three ponds on downstream temperature exemplifies the complexities of heat and temperature dynamics in beaver-impacted streams. The current study’s attempt to understand the drivers of downstream temperature at these locations thus necessitated a careful analysis of hydrogeologic and physical site characteristics. In the following discussion, I assess each measured parameter and the role it plays in regulating downstream temperature, as well as the implications of these findings.

1. Discharge

This study was conducted during the summer low-flow period in a watershed where it is not unusual for headwater tributaries to run dry. While beaver dams are associated with decreased stream velocity, they are also known to increase stream depth and stability over seasonal time scales (Westbrook et al., 2006, Nyssen et al., 2011). Modeling efforts indicate that these small differences in discharge during low-flow periods have a critical impact on stream temperature. In simulations at 4B-M-DS-1, a simulated discharge rate of 60% of the observed discharge rate resulted in 1.4 °C warmer stream temperatures over a relatively short distance (11 m).

The relationship I observed between stream discharge and temperature is in line other studies of small streams. As noted by Majerova et al. (2015), low stream discharge rates result in small
volume-to-surface area ratios, and thus greater heat exchange at the streambed and stream surface boundaries. Studies have also found that low discharge rates increase the residence time of the water, which allows heat fluxes at the streambed and stream surface to exert a strong influence over short distances (Webb and Zhang, 1999). The low discharge rates observed during the study can therefore be understood as effectively diminishing the influence of upstream temperature while amplifying the role of local heat fluxes.

These results also point to the importance of the stream rating curve and discharge measurements. A multi-year study would be preferable for calibrating the rating curve to field-collected discharge measurements; this would also allow for greater model accuracy under a wide range of discharge scenarios.

2. The stream surface boundary

The stream surface was a net source of heat at all the main study locations. At distances <100 m downstream of the 2B-M-pond and 4B-U-pond, the average net heat flux at the stream surface boundary was 68 W/m². This value is comparable to surface heat fluxes measured in previous studies of small or partially shaded streams (Łaszewski, 2005). Wider streams, such as those described in Webb and Zhang (1999) and Hebert et al. (2011), often have surface heat fluxes >130 W/m² due to more direct solar radiation. At the KRW, despite high levels of solar radiation measured at the weather station, riparian shading prevented large portion of the radiation from reaching the stream surface.

In addition to the streams, the ponds monitored in the KRW also had a positive net heat flux at the water surface boundary. Given that the ponds were mostly unshaded, this finding was expected. However, it is interesting to note that the average net flux at the pond surfaces (38 W/m²) was lower than that of the stream surfaces due to greater negative heat fluxes. Evaporation and sensible heat transfer, which were both accelerated by high pond surface temperatures, resulted in heat losses. Additionally, high rates of outgoing longwave radiation were induced by the high pond surface temperatures.
3. The streambed boundary

In stream temperature studies, the streambed portion of the heat flux budget is sometimes neglected (Hebert et al., 2011; Laszewski, 2015). While this may be justified in large, free-flowing streams, the presence of beaver dams at the sites herein greatly increased the importance of the streambed. Furthermore, upwelling from the streambed is known to be greater in magnitude, as well as responsible for a greater portion of the stream heat budget, under low-flow conditions (Mayer, 2012). At monitoring locations downstream of the dams at Site 2B and Site 4B, the streambed was responsible for an average heat flux of -34 W/m², which represented a significant portion of the stream heat budget.

The surficial site characteristics that control streambed heat flux – and in particular, hydraulic conductivity – are therefore crucial to understanding this boundary. At Sites 2B and 4B, the hydraulic conductivity values estimated from the soil analysis were similar at all measured locations (approximately $10^{-5}$ m/s). However, subsurface tracer testing found faster travel times than the soil hydraulic conductivity measurements would suggest. This discrepancy suggests variability in the deeper subsurface, as well as the presence of horizontal layering or preferential flow paths. At Site 4B, tracer flow between the pond subsurface and 4B-U-DS-1 was exceptionally fast. High rates of connectivity in this region give rise to high rates of hyporheic flow; upwelling measured at 4B-U-DS-1 were among the greatest measured at any location. The 2B site had slightly slower tracer travel times, but still a high rate of upwelling.

It is important to note that while vertical upwelling was greater at 4B-U-DS-1, both upwelling rate and overall stream temperature were more consistent at 2B-M-DS-1 and 2B-M-DS-2. This is likely related to the local hydrology and topography of the sites. For instance, pond depth, hydraulic head gradient, and downwelling rates were relatively stable at Site 2B, while these parameters varied significantly at 4B-U-pond. Interestingly, the models suggest that this consistent streambed flux at Site 2B is also more effective in cooling daily maximum stream temperatures. By contrast, the advective inputs at Site 4B – which varied hourly – were sometimes larger at night and smaller during the day, effectively cooling the stream when it was already near its minimum temperature.

Although lateral inflow was not measured in the current study, this source of hyporheic water from the streambanks might explain discrepancies between observed and predicted water
temperature at certain locations. Soil moisture is a significant control of lateral inflow (Wondzell, 2006), and the 2B Site soils appeared to be highly saturated. Furthermore, the streambanks at certain locations – 2B-M-DS-2, in particular – were quite steep, comprising a portion of the perimeter roughly equal to that of the streambed. Additionally, work by Ruehl et al. (2006) suggests that lateral inflow tends to increase during periods of low stream flow. The high streambanks, high soil moisture, and low in-stream discharge rates at certain 2B locations thus point to lateral inflows as a potential source of hyporheic flux not captured in the model.

4. Hydraulic gradient

Previous studies have found that in-stream structures may induce downwelling into otherwise gaining stream reaches (Lautz et al., 2006; Hester and Doyle, 2008). These vertical fluxes are driven by steep head differences between the dam and the downstream reach (Hester and Doyle, 2008). Downwelling was recorded in the 2B-M-pond and the 4B-U-pond, which both feature large dams and significant head gradients. At the 4B-L-pond, which was less intact and effectively flowed into the downstream channel without a change in gradient, downwelling was not recorded. Moreover, strong upwelling was recorded at both 4B-L-pond and 4B-L-DS-1. This may indicate that the hydrology of 4B-L-pond – 78 m downstream from the much larger 4B-U-pond – was influenced by gradient-driven flow from this upper pond.

High head gradients, such as those found between 4B-U-pond and the lower pond, are also known to induce long longitudinal flow pathways (Gooseff et al., 2006). However, given the shallow depth to bedrock in many regions of the KRW – including the Site 4B, where drift thickness is less than 3 m - these long flow pathways may be restricted.

5. Spatial resolution

Given the heterogenous geomorphology of the KRW, it is possible that the sensors installed at each location did not capture the full range of streambed heat flux along the study reaches. Variation in topography and sediment deposition are known to influence hyporheic exchange, with order-of-magnitude differences recorded over small spatial scales (Storey et al., 2003; Poole et al., 2006).

In the ponds, for instance, advection rate likely varied in part due to sensor placement and pond bathymetry. While sensors at 2B-M-pond and 4B-U-pond were placed 1.5 m upstream of the
dam, sensors at 4B-M-pond were placed approximately 3 m upstream from the dam on the flat pond bed. This small spatial difference may be significant; previous studies have found lower downwelling rates in deep pond pools and relatively high rates on the sloping pond glide <1.7 m from the dam (Briggs et al., 2012). This phenomena has also been described in terms of the capture zone, which describes the region of the pond (generally upstream) that experiences upwelling, and the region of the pond (generally downstream) that experiences downwelling (Feiner and Lowry, 2015). Because the extent of the capture zone may expand or contract with changing head gradients and discharge rates, it is possible that the 4B-L-pond sensors were within this range.

6. **Stream geomorphic units**

Stream morphologic features at the channel-unit scale have been found in previous studies to mediate stream temperature, particularly through control of streambed processes (Gooseff et al., 2006; Wondzell, 2006). For instance, Majerova et al. (2015) found variation by geomorphic unit (i.e. pool, riffle, and meander) in their observations of hyporheic flux. In the KRW, the relationship between geomorphic unit and heat flux or temperature regime is less clear. For instance, a similar magnitude of hyporheic upwelling was observed in pools, riffles, and runs <100 m downstream. Likewise, the deep pools all had relatively cool beds, but not significantly cooler than those of certain shallow stream segments (i.e. 2B-DS-1 and 2B-DS-3).

One geomorphic unit that did show distinct temperature dynamics was the ponds, which appeared to thermally stratify between the bed and water surface. This finding is consistent with Caissie et al. (2014), who describe this phenomenon at depths greater than 0.7 m. Stratification resulted in cooler and more stable pond bed temperatures, which in turn stabilized temperatures in the downstream channel.

7. **Aquatic habitat**

In marginal cold-water streams, low flows coupled with high temperatures may negatively impact the growth and productivity of aquatic species (Dunbar et al., 2009). Conversely, greater depth and temperature diversity in beaver dam complexes has been associated with increased trout abundance (Gard, 1961; Sigourney et al., 2006). For Rainbow Trout in the KRW, water temperatures between 10-20 °C have been found to promote growth, while temperatures above
this range (20-25.5 °C) are considered stressful (Peterson, 2012). For Brown Trout, 5-23 °C is defined as a range of growth and water temperatures 23.1-26.3 °C is considered stressful (Peterson, 2012). At Site 2B, average streambed or pond bed temperature was in the growth range for both Rainbow and Brown Trout at all measured location, and maximum daily temperature rarely exceeded this range. However, at Site 4B, average temperatures at the pond and streambed were often close to 20 °C. In the area <100 m downstream at Site 4B, streambed temperatures were slightly above 20 °C, indicating potential thermal stress to aquatic species. Additionally, at all measured Site 4B locations – including those above the beaver impoundment – maximum temperatures were in the stressful range for Rainbow Trout. The lowest maximum temperature at this site (20.6 °C) was found at the 4B-U-pond bed. These findings suggest that trout habitat in the KRW is vulnerable to thermal stress in both beaver and non-beaver impacted reaches. Beaver impoundments may provide thermal refuge and cool downstream temperatures in some cases, but in other cases may induce stressful temperatures at distances <100 m downstream of the impoundment.

8. Modeling

The models used in the current study (MNSTREM and VFLUX) are well-developed and have been applied successfully in numerous other stream studies. Furthermore, multiple studies that apply MNSTREM have taken place in the same geomorphic region of northern Minnesota examined herein (Sinokrot et al., 1995; Herb et al., 2008). Although my application of the MNSTREM model to dam-impacted streams is less straightforward, it is not without precedent (Sinokrot et al., 1995). It is also important to note that the modeling results seem reasonably well-matched to the field experiment, and that both the field experiment and models agree with basic hydrologic theories. However, given the high heterogeneity of Minnesota streams, these models should be primarily used to assess and compare temperature dynamics in the local stream environment. In other words, the results are site-specific do not fully explain thermal dynamics in other stream locations.

The modeling efforts in this study were somewhat inhibited – particularly at Site 2B – by equipment failures at in-stream monitoring locations. A higher in-stream monitoring resolution would be useful in calibrating the models, and a second field season would allow for model validation. To expand upon this research, I would recommend 1) depth and streambed sensors at distances >100 m downstream, 2) a more precise quantification of streambank and riparian
shading, 3) temperature sensors at the water surface for all locations >0.5 m deep, and 4) stream monitoring both before and after beaver dam removal.

Through two validation periods, I demonstrated that the models in this study may be applied during time frames beyond the calibration period. However, as illustrated through the August validation period, data collected on an hourly time step are necessary to achieve accurate model results. This is particularly true of parameters that fluctuate strongly on diel timescales, such as solar radiation and air temperature.

9. Future Work

Previous studies of the hyporheic zone have provided insight into how water masses move between the stream and shallow subsurface. However, there is a dearth of research on how these flows operate in and around beaver impoundments, particularly in watersheds with high head dams and large ponds. Higher-resolution monitoring across the breadth of beaver pond bathymetric features would be useful in understanding the spatial dimensions of upwelling or downwelling seepage.

Stream shading studies in beaver-impacted watersheds are also uncommon. Beavers activity significantly alters riparian vegetation, leading to complex shading patterns along the streambank. Further examination of how beaver impoundments alter stream vegetation, as well as how these variably vegetated riparian zones impact stream temperature, would help to illuminate the stream heat dynamics of these environments.

In recent years, a significant amount of research on beaver-altered streams has taken place in mountainous regions of the Western U.S. I contend that applying the research ideas herein to watersheds in low-gradient streams, and streams in the Great Lakes region more specifically, would provide insight into the stream thermal regimes of these under-studied environments.

VIII. CONCLUSION

Beavers impact stream morphology in a variety of ways, primarily by altering flow pathways and water storage distribution between surface and subsurface reservoirs. These beaver dam-induced changes to ponds, streams, and adjacent subsurface zones are accompanied by changes to the thermal regime. In this study of low-gradient tributaries in the northern Great Lakes region, beaver dams were found to alter stream geomorphic template and temperature most strongly at
distances <100 m downstream. By 300 m downstream, impacts to stream temperature were undetectable.

Beaver impoundments are known to increase the spatial complexity of stream environments. In heterogeneous post-glacial landscapes, beaver dam impacts are even more variable. Highly site-specific thermal trends were observed downstream of impoundments in the study tributaries, particularly at the streambed. The magnitude of advective and conductive heat fluxes at the streambed varied between sites and on spatial scales of <10 m. These findings indicate that high-resolution spatial measurements and characterization of the surficial geology are crucial to understanding the thermal regime of these landscapes.

My findings further suggest that stream discharge rate is a strong control of stream temperature, particularly in small headwater tributaries. During the low-flow period, low discharge rates were observed to exacerbate thermal warming trends. In regions where cold-water stream habitat is at risk, analysis of streamflow volume and timing may therefore be important to understanding and protecting these environments.
IX. BIBLIOGRAPHY


81

Tappe, D. T. (1942). The status of beavers in California. State of California, Department of Natural Resources, Division of Fish and Game.


X. APPENDIX

Table 14. Coordinate location and elevation of all monitoring locations at Site 2B and Site 4B.

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Table 15. Site 2B maximum daily temperature measured upstream of the pond, at the pond bed and surface, and at seven in-stream locations downstream of the pond.

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### Table 16. Site 4B maximum daily temperature measured upstream of the pond, at the pond bed and surface, and at seven in-stream locations downstream of the pond.

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Table 17. Equipment specifications

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<th>Equipment</th>
<th>Range</th>
<th>Accuracy</th>
<th>Resolution</th>
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<tr>
<td>Solar Radiation (Silicon Pyranometer)</td>
<td>0 to 1280 W/m²</td>
<td>±10 W/m² or ±5%, whichever is greater in sunlight; additional error ±0.38 W/m² /°C from 25°C</td>
<td>1.25 W/m²</td>
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<tr>
<td>Smart Sensor (S-LIB-M003)</td>
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<tr>
<td>12-bit Temperature/Relative Humidity (2m cable)</td>
<td>Temp: -40°C to 75°C; RH: 0-100% RH at -40°C to 75°C</td>
<td>Temp: +/- 0.21°C from 0°C to 50°C; RH: +/- 2.5% from 10% to 90% RH (typical); below 10% and above 90% ±5% typical</td>
<td>0.02°C at 25°C; RH: 0.1% RH</td>
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<tr>
<td>Smart Sensor (S-THB-M002)</td>
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<tr>
<td>Wind Speed Smart Sensor (S-WSB-M003)</td>
<td>0 to 76 m/s</td>
<td>±1.1m/s or ±4% of reading, whichever is greater</td>
<td>0.5 m/s</td>
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<tr>
<td>HOBO Water Temp Pro v2 (U22-001)</td>
<td>-40°C to 70°C in air; max sustained temp of 50°C in water</td>
<td>±0.21°C from 0°C to 50°C</td>
<td>0.02°C at 25°C</td>
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<tr>
<td>HOBO 30-Foot Depth Water Level Logger (U20-01-001)</td>
<td>Temp: -20°C to 50°C; Pressure: 0 to 207 kPa; Water level: 0 to 9 m of water depth at sea level</td>
<td>Temp: ±0.44°C from 0°C to 50°C; Pressure: ±0.3% FS, 0.62 kPa maximum error; Water level: Typical error: ±0.05% FS, 0.5 cm water Max error: ±0.1% FS, 1.0 cm water</td>
<td>Temp: 0.10°C at 25°C; Pressure: &lt;0.02 kPa; Water level: 0.21 cm water</td>
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<td>HOBO Water Level (13 ft Data Logger (U20-001-04)</td>
<td>Temp: -20°C to 50°C; Pressure: 0 to 145 kPa; Water level: 0 to 4 m of water depth at sea level</td>
<td>Temp: ±0.44°C from 0°C to 50°C; Pressure: ±0.3% FS, 0.43 kPa maximum error; Water level: Typical error: ±0.1% FS, 0.4 cm water Max error: ±0.2% FS, 0.8 cm water</td>
<td>Temp: 0.10°C at 25°C; Pressure: &lt;0.014 kPa; Water level: 0.14 cm water</td>
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<td>Thermochron DS1921H F5# iButton</td>
<td>15 °C – 46 °C</td>
<td>±1 °C</td>
<td>0.125 °C</td>
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<tr>
<td>Seapoint Rhodamine Fluorometer</td>
<td>Excitation: 540 nm peak; Emission: 610 nm peak; Temp: 0°C to 65°C</td>
<td>Nominal range and sensitivity at a given gain setting</td>
<td>Min. Detectable: 0.02 µg/l; Sensitivity: 0.033, 0.1, 0.33, and 1 V/(µg/l) are possible</td>
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<tr>
<td>TD-700 Laboratory Fluorometer</td>
<td>Detection: 300 - 650 nm; Temp: 5 - 40°C</td>
<td>Dependent on user calibration</td>
<td>Sensitivity: less than 20 picograms</td>
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